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VALLEY DEPOSITS IMMEDIATELY WEST OF THE CHANNELED SCABLAND

J HARLEN BRETZ
University of Chicago

ABSTRACT

If a "Spokane Flood" ever occurred, and if the valley system of eastern and central Washington was already in existence, these valleys must have become channels and abnormal deposits of great magnitude must have been formed. This paper calls attention to such deposits along the western margin of the scabland—both in Columbia Valley, a main discharge route, and in Yakima Valley, a capacious tributary valley entering from an area unaffected by the flood. Three extraordinary situations are given especial attention: the mouth of Moses Coulee, the Columbia Valley deposits near Trinidad, and the truly astounding features in lower Yakima Valley. The conclusion is reached that nothing short of a catastrophic flood can explain these features. Skepticism can be justified only by the construction of workable alternatives.

The loess-mantled basalt plateau of eastern Washington possesses a system of linear anastomosing denuded tracts, the channeled scabland, in parts of which the basalt has been trenched in most unusual fashion. The writer has interpreted these linear tracts as abandoned river channels, rather than river valleys. From features of the anastomosing pattern, it was concluded also that most of the channels were contemporaneously occupied. These two ideas required the assumption of a volume of stream water greatly exceeding anything known elsewhere. Other data seemed to indicate that the great rivers were born suddenly, operated for a very brief time, and then abruptly ran dry. The term "flood" has been used to express this conception.

Doubt of the correctness of the writer's interpretation was early

expressed, and several alternatives were suggested for the various peculiar features and combinations of features of the channeled scabland. In a paper considering these alternatives,[1] the writer noted that the flood hypothesis was not to stand or fall on data then in hand; that it could be critically tested by further field study. Certain relationships in unexamined areas must exist if it be correct and could not exist if it were incorrect. This paper deals with these relationships along the western margin of the main tract of channeled scabland. The areas considered are the Columbia Valley from Wenatchee River to Yakima River and the lower part of Yakima Valley.

COLUMBIA VALLEY FROM WENATCHEE TO THE MOUTH OF MOSES COULEE

In the paper above noted, it was stated that if the flood hypothesis is correct, Columbia Valley above the mouth of Moses Coulee must contain one of two types of record: (*a*) that of vigorous southward flow of an enormous glacial river whose surface descended from the edge of the ice to an altitude, at the junction, of from 600 to 700 feet above the present valley bottom, or (*b*) that of a backwater ponding produced by the great scabland rivers entering Columbia Valley at, and south of, Moses Coulee. If, on investigation, "neither of these exist, the hypothesis is fundamentally wrong."

The writer, after study of this part of the valley, concludes that neither record is present.

Obviously, the flood hypothesis must now be abandoned or the ultimatum must be retracted. If the latter alternative be followed, some special explanation must be devised for the absence of a flood record up the Columbia from the mouth of Moses Coulee.

No ingenuity is required to construct this special explanation on the basis of field evidence unknown when the ultimatum was made. Wenatchee River, draining from the high Cascades, enters the Columbia about 15 miles upstream from the mouth of Moses Coulee, and Columbia Valley between these two tributaries carries an unmistakable record of glaciation. Moraine topography extends down

[1] "Alternative Hypotheses for Channeled Scabland," *Jour. Geol.*, Vol. XXXVI (1928), pp. 193–223, 312–41.

the Columbia virtually to the entrance of the great coulee. Except that it is pre-Wisconsin, the age of this glaciation is unknown. To save the flood hypothesis, it is only necessary to assume that the glaciation was contemporaneous with, or subsequent to, the flood. The critical reader may object that advantage is being unfairly taken of gaps in our knowledge. The writer's reply is to counsel patience until the entire paper has been read.

The major features of Columbia Valley between Wenatchee and the mouth of Moses Coulee are determined by two formations, the Swauk shale and sandstone beneath the resistant Columbia River basalt. The dip is eastward, in the direction the river flows but more steeply. Thus the river has a wide valley in the Swauk formation for about 10 miles downstream from Wenatchee, overlooked on the north and south by converging basalt escarpments from 1,700 to 1,800 feet high (Fig. 1). About 5 miles above the mouth of Moses Coulee, the base of these escarpments comes down to river level and the valley thence downstream for many miles is narrow and deep, with precipitous walls of basalt. Rock Island Rapids is a mile or so within this narrow valley.

Rising westward with the dip, the contact of basalt on sandstone and shale, 7 or 8 miles west of Rock Island Rapids, comes up about 1,800 feet above river level. Here the steep cliffs of Columbia basalt above the contact are 2,000 feet high and the Swauk outcrop below, 4 or 5 miles wide, is marked by landslide topography, particularly on the south side of the river. There is no possibility of mistaking this landslide topography for moraine. But over the lower 500 or 600 feet of the Swauk outcrop area, foreign boulders and cobbles are not uncommon, though none were found with glacial markings. In the vicinity of Malaga and Mud Lake, from 7 to 8 miles below Wenatchee and on the south side of the river, good moraine topography is superposed on the lowest of the landslip features. The local relief of the moraine deposit itself is about 30 feet. As a mantle or thick smear, it extends from a flattish morainic terrace for 125 feet up on the landslide forms, culminating in a definite lateral moraine ridge 300 feet above the river. On the north side of the river, opposite Malaga, the morainic topography begins about 200 feet above the river and rises northward through 400

FIG. 1.—Malaga Quadrangle, Washington, United States Geological Survey

feet of altitude to a plain of outwash sand and gravel at from 1,200 to 1,250 feet A.T., 300 feet higher than the moraine on the south side and approximately as high as the foreign cobbles and boulders on that side. The outwash plain here is about a mile wide and slopes gently northward away from the moraine toward the lower exposed slopes of the Swauk. Some of the morainic ridges and associated linear undrained hollows are well shown on the Malaga sheet of the United States Geological Survey. They extend along the slope somewhat like landslip forms, but can hardly be confused with such. The maximum relief between a ridge and its associated sag on the uphill side is 125 feet.

The morainic topography downstream in the basalt-walled valley occurs only on the floor. The local relief between hillocks and adjacent undrained sags is about 30 feet. A stratified valley fill beneath the moraine consists in part of well-varved clay. The morainic débris is hardly 50 feet thick where shown in section just south of Rock Island Rapids. From this place the general moraine surface rises about a hundred feet in the next 2 miles downstream and there, without terminal ridging, ends abruptly against, or apparently against, an enormous deposit of débris at the mouth of Moses Coulee, a deposit higher than the moraine and of quite different character.

Though the upper limit of morainic material in this part of Columbia Valley is not sharply marked, present data indicate that it descends about 450 feet in 10 miles. The river in that distance, including the rapids, descends less than 50 feet.

The composition of these morainic forms near Malaga is not the physically heterogeneous stony and bouldery clay commonly termed till. Instead, the material is gravel, the fragments worn by running water instead of glacial ice and the finer constituents simply filling voids among the pebbles, cobbles, and boulders. Some sections show a poor stratification and sorting. It looks as though glacial ice had encountered great quantities of river gravel in this stretch and had re-worked it to make a morainic material largely devoid of clay and without evidence of glacial abrasion. Similar very gravelly moraine has been seen in the Rocky Mountain Trench at Polson and in the Purcell Trench north of Lake Coeur d'Alene, both places

affording the possibility of glacial re-working of large stream-gravel deposits.

The content of basalt in this gravelly waste changes remarkably along the valley length considered. Near Mud Lake and Malaga, basalt constitutes from 30 to 35 per cent of the material. Three miles farther downstream, in the lee of some basalt hills that appear to be huge slide blocks out in the valley, the moraine hillocks are littered with large fragments of that rock, and 2 miles still farther downstream (at the rapids) the morainic débris is 85 per cent basalt,

Whether this ice came down the Wenatchee or Columbia valleys. or both, is not yet known. The first alternative seems most likely. Whether it was contemporaneous with the great scabland-making episode or was later is not yet determined. It seems, however, to provide a logical explanation for the absence of any record of Spokane Flood waters above the mouth of Moses Coulee.

COLUMBIA VALLEY AT THE MOUTH OF MOSES COULEE

A huge flat-topped mound of coarse detritus stands out in the middle of Columbia Valley at the junction of the great dry valley of Moses Coulee, the westernmost of the scabland river channels. The Columbia here is crowded against the west wall of the valley and confined to a narrow trench 350 feet below the highest part of the mound.

The mound occupies two-thirds of the valley width. Its summit area is triangular in outline, the highest part constituting an acute angle pointing up the Columbia Valley (Fig. 2). From this summit, its surface descends 50 feet per mile toward the south and southeast to the ravine draining out of Moses Coulee. But both sides of the acute angle are steep bluffs, the northeastern one descending about 50 feet to the moraine and the western one descending 350 feet to the Columbia. The western bluff is certainly river-cut. The northeastern bluff is either river-cut or an ice-contact slope related to the glacier that came this far down the Columbia. The gentle southern and southeastern slope is constructional. The mound summit is 350 feet above the river, 100 feet above the lowest place between it and the rock wall to the northeast, and about 150 feet above the ravine to the southeast.

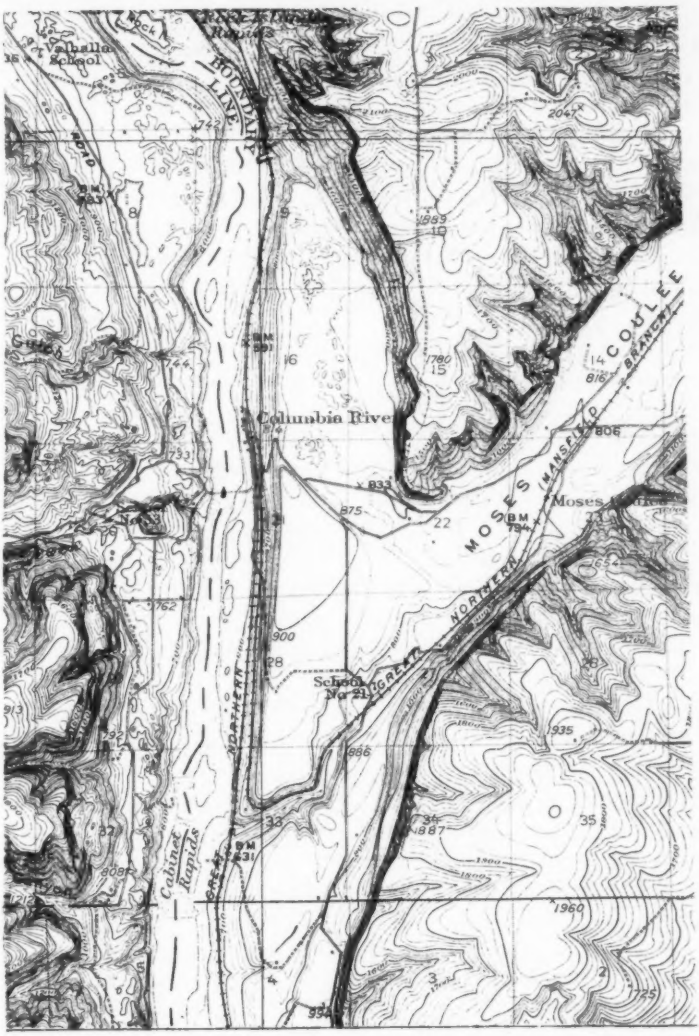

FIG. 2.—Junction of Columbia Valley and Moses Coulee; part of Malaga Quadrangle, Washington, United States Geological Survey.

There are no rock outcrops in the 350-foot western bluff or in the ravine from Moses Coulee. The material is apparently coarse gravel and boulders from top to bottom. Wells farther up Moses Coulee show corresponding depths of gravel below a floor approximately as high as the mound summit.

This mounded deposit, standing nearly in the middle of Columbia Valley at the mouth of Moses Coulee and undefended by any rock knob, is far higher above river level than any other deposit out in the valley for many miles upstream. It is alone as a mid-valley mound of débris; all other deposits in such position are terrace forms.

A plausible suggestion is that the location of the deposit is in some way related to the mouthing of Moses Coulee. Wenatchee Valley is comparable to Moses Coulee in width and depth, but there is no such deposit in the master valley at its mouth. Perhaps, therefore, different events in these two tributary valleys are indicated. Since Moses Coulee drains only from the basalt plateau, débris derived from it should be predominantly basaltic (though glacial ice on the north edge of the plateau might contribute some non-basalt). Columbia Valley receives drainage from a great variety of rocks, and its débris should be exceedingly varied. This criterion seems worth applying to the great mounded deposit.

Columbia River gravel now in transit rarely shows basalt higher than 50 per cent. This is true well below Moses Coulee, in the narrow basalt-floored and basalt-walled valley. No nearby tributaries from non-basalt areas are responsible for this proportion; it is due to non-basaltic débris brought down the Columbia past the mouth of Moses Coulee. Gravel of older terraces of Columbia Valley about Wenatchee and upstream show perhaps one-fourth basalt. The same proportions hold for the younger terraces, probably Wisconsin in age.

But the great mound is from 90 to 99 per cent basalt! This startling fact indicates clearly that the deposit consists almost wholly of débris from Moses Coulee, and that Columbia River has contributed little or nothing to it. If one considers again its remarkable proportions and its location out in mid-valley, it is difficult to escape the conclusion that Columbia Valley has been almost

completely blocked here at some time by very coarse débris from a now dry tributary canyon. It is not difficult to visualize a complete temporary blocking and to devise criteria by which that idea might be tested. If the Columbia's trench along the west side of the mound has been cut 350 feet deep in an originally continuous fill of basaltic débris, there might be remnants still lingering on the west side of the river, remnants identifiable by altitude, composition, and structure.

Two minor tributary valleys from the west enter the Columbia opposite the great mound. Both contain valley-mouth deposits of gravel which have obviously caused relocation of the stream courses on buried rock shoulders. Both streams enter and cascade through short, steep-walled gorges within half a mile of the Columbia. In both, the deposit is limited to the valley mouth; it does not extend back up the tributary.

The deposit in the valley of the southern tributary, Colocham Creek, reaches 1,000 feet A.T., 75 feet higher than the summit of the mound across the river. That in Dry Gulch is about 100 feet above the mound summit. In both, the débris is from 90 to 99 per cent basalt. This would be expected in tributaries draining, as these do, entirely from basalt country. But in both the foresets dip back into the mouths of the tributary valleys, out of the Columbia and toward its western wall. These features seem to require some special explanation; they should not exist in Columbia River gravel deposits.

A further anomalous feature is found in the shapes of the gravelly fragments. In contrast with Columbia River gravel, which is well worn, most of this débris is sharply angular. Extraordinarily thin blades and long jagged slivers of basalt are very common. Rounded surfaces are present, but most fragments possessing them also have some sharp, hackly, fractured surfaces in addition. Such pebbles show clearly that they were originally rounded, were then broken by sharp blows, and promptly thereafter came to rest. The sharpness of fractured outlines of both broken-round and wholly angular fragments absolutely debars the idea of ordinary travel along any watercourse after the breaking.

The deposit in the mouth of Dry Gulch extends as a terrace-like

form about a mile farther north along the Columbia Valley wall, its surface descending nearly 100 feet in that distance. This is as difficult to explain as a Columbia Valley feature as are the items already listed. It is perfectly in harmony with the picture of a great deposit from Moses Coulee blocking Columbia Valley. It needs only foreset bedding with dip up the Columbia to make it complete. Unfortunately, no exposures of its structure were found. Higher than this terrace are fragmentary remains of gravel deposits dominantly non-basalt. These seem to record much earlier Columbia River gravel, antedating even the last 700 feet of erosion into basalt, all of which occurred before the basaltic gravel deposit was made.

If one again takes stock of the data to see what other credit items for the Spokane Flood hypothesis should exist here, he finds that erosional forms apparently recording such an extraordinary episode are reported for Moses Coulee alone. He asks logically for scabland features on the west wall of Columbia Valley here, produced by the great flood which is hypothecated for the deposits. That scabland exists. It lies between the two tributary valley mouths and below about 1,250 feet A.T. A flattish basalt ledge here, a mile long and about half a mile wide, has been swept clean of mantle rock,[1] and the bare basalt has been notably roughened into crags and buttes and rock basins that show well on the Malaga sheet. A steep cliff rises back of this scabland ledge, a cliff which clearly is a truncated shoulder, formerly rounded and much less steep. Except for this cliff, virtually all slopes above 1,250 feet have a graded waste cover and are marked by normal drainage forms.

Moses Coulee has been pictured in the writer's earlier papers on scabland as a preglacial drainage line heading about 50 miles to the northward on the plateau and amazingly eroded later by glacial waters. A great ragged, dry cataract, 400 feet high and 1 mile wide, crosses the coulee about 20 miles above its mouth; huge bar-shaped gravel deposits occur throughout a vertical range of 300 to 350 feet on the greatly scoured lower slopes; and a major tributary, Douglas Coulee, entering the main canyon about 16 miles above the mouth, hangs 350 feet above the gravel floor and 550 feet above

[1] Deposition of dust and disruption of basalt have started a new mantle.

the rock floor at its point of entrance. Many minor hanging valleys are shown on the Malaga sheet. Much of this hanging condition of tributaries, which surely were adjusted to the preglacial valley floor, appears to be due to sapping of the coulee walls and consequent widening by glacial waters.[1]

The relation of this Moses Coulee discharge to the valley glacier that once reached down the Columbia to the coulee mouth needs a few more words. So far as altitudes and topographic relations are concerned, one might picture the great mound as outwash material from this glacier, the accumulation having been built up from 50 to 100 feet higher than the morainic topography immediately behind it. The sag northeast of the mound summit might be considered a spillway from this ice. But by such an origin the great mounded deposit would have a large proportion, if not a dominant amount, of non-basalt in its débris and its imbricated and foreset structures would be directed down, not across, the Columbia.

It can hardly be argued that a Columbia Valley glacier extending down to this junction would yield so little water that Moses Coulee's normal share of the ice sheet's drainage across the plateau would have given it complete dominance over the Columbia discharge. There must be added to the coulee discharge a very great supply from some extraordinary source.

The history of Moses Coulee, therefore, appears to include an episode of tremendous discharge of glacial water down through a canyoned tributary to a Columbia Valley as deep then as now, remarkable cataract development and recession during the life of the enormous discharge, remarkable widening to make of the pre-existing valley an adequate channel for the glacial river, bar-building on a scale known elsewhere only in the channeled scabland of the Columbia Plateau, and the making of a mammoth bar completely

[1] The 400-foot cataract, however, records at least that amount of deepening immediately downstream from it. The cataract probably became higher as it receded. It began, therefore, somewhere near the junction of Douglas Coulee, as a relatively minor affair. Francis Canyon, shown on the Malaga sheet, does not hang above the gravel fill of Moses Coulee, but a dug well in mid-coulee a mile distant penetrated 300 feet of gravel without finding bedrock. Francis Canyon, therefore, does hang at least about 300 feet above the true bottom of Moses Coulee. Its gradient, projected to mid-coulee, would not be 100 feet below the surface of the gravel fill.

across the Columbia Valley at its mouth. That a long series of Pleistocene episodes was not involved in making these great changes (though other Pleistocene events have undoubtedly occurred here) is argued from the following items:

1. Douglas Coulee, which never received glacial water of any great amount, is a normal valley for the region and hangs far above the main coulee. Had interglacial stream erosion been an important factor in the deepening and widening of Moses Coulee, this tributary would not hang. The erosion in Moses Coulee, therefore, was done by glacial water.

2. Hanging tributaries, if by any chance the result of interglacial stream erosion along main drainage lines, would exist in Douglas Coulee, Rock Island Creek, and other canyons of the region.

3. A long-continued aggradation in Columbia Valley at the mouth of Moses Coulee would have mingled débris of the master stream throughout. The basaltic content then would have been about 50 per cent instead of from 90 to 99 per cent.

4. Foresets in the great deposit and in its remnants on the west side of the Columbia would not dip westward, as they do, across the master valley if the Moses Coulee glacial river had not completely dominated the Columbia at that time. Imbrication of the flattish cobbles also indicates that the transporting current came from Moses Coulee, not Columbia Valley.

5. Virtually the only identifiable Columbia River débris in the remnants west of the river are the few rounded pebbles and the somewhat more numerous broken rounds. These are about 50 per cent non-basalt. The sharply angular material is essentially all basalt. No normal stream work produces such percussion-broken chips and slivers, or breaks up its rolled forms in the fashion or to the extent that these broken rounds record. The rounds and broken rounds were probably derived from Columbia River gravel already on the valley floor when the flood occurred.

COLUMBIA VALLEY BETWEEN MOSES COULEE AND THE POTHOLES CATARACT

The great bar at the mouth of Moses Coulee was built largely on the northern side of the glacial torrent. Its gentle southern

aggradational slope is the northern part of the glacial river channel floor at the close of the episode. The gravel, therefore, was thrown back up the Columbia 100 feet higher than it was left in mid-channel. The southern part of this subfluvial gravel deposit lies south of the ravine subsequently trenched in the middle of the old channel. Just south of the ravine, it rises from mid-valley to the base of the eastern cliffs through a vertical range of from 100 to 125 feet. A

FIG. 3.—Vicinity of Trinidad, Washington. Parts of Colockum Pass, Washington, and Quincy, Washington, Quadrangles, United States Geological Survey.

terrace in this slope looks a bit confusing to this interpretation and will be dealt with later.

Four miles below the mouth of Moses Coulee, the Columbia turns sharply to the east for about 3 miles and then, at Trinidad, resumes its southward course (see Fig. 3). Northeast of the river, on the inside of the first curve noted, a thick gravel deposit extends from river level up to 1,100 feet A.T., a height of 550 feet. Nowhere does it have terrace forms. That it is a bar deposit still retaining the original form, except for obvious gullying, is shown in both topog-

raphy and structure. Its structure is one of the most remarkable features to be reported in this paper.

The Colockum Pass map of the United States Geological Survey shows the highest part of this great gravel deposit as a compound lobate form depending diagonally down valley from the sharp turn in the basalt wall north of, and above, the gravel. The highest place in the deposit is a broad low mound half a mile long, lying nearly half a mile out in the valley and separated from the base of the rock wall to the north by a shallow fosse. On the north side of this mound, the gravel is foreset up the Columbia. The location is ideal for an eddy in a great stream filling the Columbia Valley from side to side at this altitude, and this low mound and its structure can be explained only by the conception of an eddy.

But the truly remarkable structure of this bar deposit is found in the riverward slope. The Great Northern Railroad has a large pit here, the cut bank of which is 200 feet or so high. Throughout this cut, the bedding dips *down to the west*, directly *up* the Columbia and parallel with the upvalley slope of the bar itself. In several places, there are lenses of short foresets in these dipping beds and the foresets dip *east*, *down* the Columbia, and against the dip of the gravel as a whole. They dip in toward the body of the deposit.

The writer has studied the structure of stream bars at every opportunity since he recognized these great scabland gravel mounds as genetically the same. He has found that flood-time bars, which become obliterated in large part before another flood and hence are essentially new constructions of each flood, commonly show stratification on all slopes parallel with those slopes. These bars, if properly placed, grow on the upstream side as well as laterally and downstream. Accretion on the upstream face thus produces inclined bedding which dips upstream, descending against the direction of the débris-bearing current. Unlike foresets, the material of such inclined beds has been moved up the slope.

The bar west of Trinidad, by this interpretation, was built by a current coming down the Columbia Valley. The predominant basaltic composition of all this banked-up gravel from Moses Coulee down to Trinidad indicates derivation from the tributary coulee, not the main Columbia Valley. Its change from very cobbly and

bouldery gravel to material of railway ballast grades is perfectly logical. But to build this great deposit, a river must be sweeping over the total range in altitude simultaneously. The central current must be as deep as the bar is thick. If we measure from the base of the pit to the summit of the bar, that stream was 350 feet deep when the bar was growing on its upstream face. Up over that slope, the current dragged débris and added it until an equilibrium size for the eddy, or slacker place, in the current was attained.

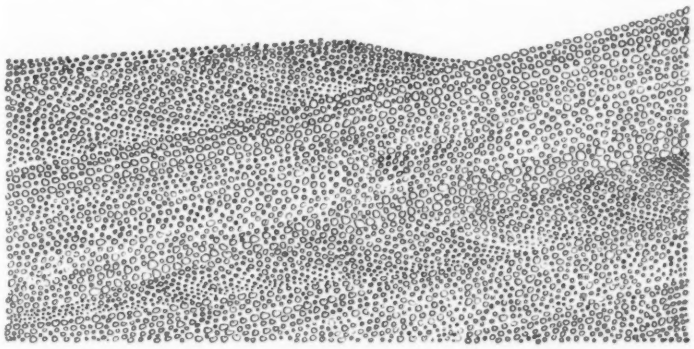

FIG. 4.—Diagram to show accretion on the upstream slope of a river bar and the development locally of current foresets dipping with the current but against the general structure and slope.

Locally, while sweeping gravel up this subfluvial slope, the current made downstream foresets which show by their dip, their gradation of material, and their decreasing amount of dip toward the lower ends of laminae that the current undeniably was moving along and up this slope (see Fig. 4). The central current was vigorous enough, after bar accumulation had filled in the slacker places, to keep the main channel from filling. Is there any other explanation for such structure and topography of a river-gravel deposit?

Another very instructive section exists in a second large pit, happily located at the downstream terminus of this bar deposit. Only coarse sand and fine gravel are exposed, well stratified and sorted. Here also the beds are dipping. But here the dip throughout is inclined toward the *east*, *down* the Columbia Valley. There

are no current foresets in this; the dipping beds are more like long deltaic foresets. This section is in the lee face of the great side-channel bar, structurally, topographically, and texturally. It is not possible to make a fan or a delta of the deposit. Nor is it possible to make of it a series of accumulations during several distinct episodes of Pleistocene history.

The true vertical proportions of the bar, and therefore the maximum depth of the gigantic river debouching from Moses Coulee, are somewhat in doubt. The base of both pit sections is about 250 feet above the river. To include this 250 feet in the bar would hardly be justifiable in the absence of evidence below track level of the same character as that above. The Columbia undoubtedly has cut into the flood-gravel deposits. How much intrenchment has occurred here?

A bit of evidence which seems to bear on this question is found in West Bar, a terrace-like form from 150 to 250 feet above the river on the inside of the second curve noted and extending 4 miles down the Columbia below and across from the bar just described. The surface of West Bar ranges through 100 feet of altitude and is remarkably diversified with large hillocks of gravel. It completely blocks the mouth of Tekison Creek Valley from the west and the tributary has been forced to turn abruptly to the south and to go around the lower end of the deposit to reach the Columbia. Seen from viewpoints along the highway east of the river, the surface of West Bar seems to be marked by great current ripples. Since the top of the bar is at about the same altitude as the base of the pit sections, it seems fair to consider West Bar a part of the actual bed of the enormous glacial river at its closing stage. In such case the Columbia has cut about 200 feet here since the episode closed. The maximum depth of the river near Trinidad, by this view, was at least 350 feet. Since the curve in the river channel was not shared here by the western rock wall, the glacial current swung hard against the base of that wall and there was no opportunity for building a correlative of the thick side-channel bar that lies inside the first curve.

An erosional record of the glacial flooding in Columbia Valley near Trinidad should accompany the depositional record. Scabland

forms, being essentially channel features, are better shown on flattish tracts than on cliff faces. We must look, therefore, for broader ledges along the Columbia at altitudes sufficiently low to have been overrun by the flood. Such ledges exist on both sides of the valley south of Trinidad. Those on the west flank the mouth of Tekison Creek Valley near the south end of West Bar. Here well-developed scabland extends up to 1,250 or 1,300 feet, from 750 to 800 feet above the Columbia. It consists of buttes, basins, and channels etched out of a bare rock terrace (a very thin soil is forming in some places) and a definite scarp rising from 100 to 150 feet higher to the west of the scoured bench. The scarp has a slope of about 27 degrees and truncates an older slope of about 12 degrees. Conspicuous remnants of the older slope, completely mantled with waste, descend toward the top of the steep scarp.

On the east side of the river is a very marked ledge of basalt, 1 mile wide in places, extending 16 miles along the river and rising from 900 feet at the south end to 1,375 feet A.T. at the north end. Most of it is shown on the Quincy sheet of the United States Geological Survey. The writer has previously described the features of this ledge, and they will be but briefly summarized here. The surface bears strongly expressed scabland, associated with small bar deposits. But this scabland is significantly limited to that portion of the ledge which lies below about 1,250 feet A.T. The northern and higher part is totally lacking in the bare-rock surfaces, the buttes, rock basins, channels, and bar deposits of scabland. It also lacks the very steep cliff overlooking it on the east. North of the scabland-marked part of the ledge, the cliff has gentle slopes and, like the ledge, is mantled completely with soil. Whether one accepts the flood hypothesis for the origin of scabland or prefers some alternative, he must admit that subsequent to the erosion of Columbia Valley and the development of this marked ledge, some pronounced scouring has occurred over that part of the ledge below 1,250 to 1,300 feet and has not occurred above that altitude. The great bar deposit west of Trinidad is the highest known in this region. Though it falls 150 feet short of reaching to the scabland upper limit, its summit is higher than most of the scabland on the broad ledge. Furthermore, a spur of basalt $\frac{1}{2}$ mile north and a little west of the

1,100-foot summit of the bar has been eroded into a jagged crest line at least 100 feet above the bar top. Closely associated with the largest notch in the spur is a mounded heap of basaltic débris thrown over into the ravine and interrupting its normal form. It seems probable that if one accepts the flood hypothesis for the bar, he must add the scabland-making to the episode and must therefore think of a river 500 feet deep.

CRATER CATARACT, NEAR TRINIDAD

The western part of the Quincy sheet shows three abandoned cataracts once used by glacial water escaping from Quincy Basin, in the plateau scabland, and entering the Columbia. In each the water above the brink reached essentially the same upper limit, from 1,275 to 1,300 feet A.T., and they must be considered as contemporaneous in action. This altitude is also that of the highest scabland on the broad ledge, and therefore no cataract spill could have occurred when the highest part of the ledge was scoured. Similarly, the scabland near the mouths of Colocham Creek and Tekison Creek is almost as high as the river surfaces just above the cataracts. The cataracts, therefore, must be a later development, and two episodes of scabland-making by glacial waters seem to be recorded. Several critics have urged the extreme improbability of the view that all scabland channels were contemporaneously occupied. Here is a critical place to test the idea.

Two of the cataracts (Frenchman Springs and The Potholes) poured directly over the wall of Columbia Valley on to the broad ledge, the deepest plunge pools (now dry) being 200 feet below the surface of the ledge. The northernmost fall (Crater Cataract) entered a tributary of the Columbia (Willow Creek Draw) about 2 miles upstream from the Columbia. It plunged from 1,200 to 1,000 feet A.T., and a flat-surfaced gravel fill at its base reaches thence nearly 3 miles to Trinidad, descending 50 feet in that distance and terminating in a bluff 400 feet high, now being undercut by the Columbia. The material of the fill is well exposed in numerous sections. It is composed of dominantly basaltic débris, little worn, ranging from sand to boulders but mostly a pebbly gravel. Its stratification is essentially horizontal. No foreset strata were found.

The gulch of Willow Springs Creek trenches buried rock spurs in a few places, but a creek valley as deep as at present existed here before the fill was made.

The brink of this flat valley fill is 250 feet higher than the scarped edge of West Bar 1 mile distant and 150 feet higher than the valley-wall contact of this bar, 2 miles distant. Yet the brink is 150 feet lower than the summit of the side-channel bar west of Trinidad. The topography and structure seem to rule it out of the bar category, yet its range in altitude includes that of the remarkable 200-foot section with bar structure only 2 miles farther up the Columbia. Its structure and cliffed terminus do not indicate a delta front; its gently descending gradient seems to require that it once extended farther out into Columbia Valley.

If this be true, the lower corrugated surface of West Bar, here interpreted as channel floor of the bar-building episode, must be younger. Yet the argument which ties the great bars into the episode of highest scabland-making requires Crater Cataract to function later and consequently the flat-topped Willow Springs fill to be the younger.

There are almost no terraces in Columbia Valley or in the mouths of tributary valleys for many miles upstream or downstream to use in determining age relations of this Willow Springs deposit to the bars. The only terracing of bar gravel known in the region occurs about 1 mile below the mouth of Moses Coulee. Here a scarp from 75 to 100 feet high interrupts the aggradational profile of the gravel deposit, the flattish tract below being about 850 feet A.T., 100 feet lower than the brink of the Willow Springs deposit. This scarp seems to be related to the Moses Coulee River, rather than to the Columbia, and is probably a record of channel-shifting beneath the stream surface in which the entire deposit was made.

The nearly complete absence of terraces on the bar profiles strongly suggests that the Willow Springs deposit is a product of special local conditions, and that it really belongs to the episode of tremendous discharge of glacial water across the scablands. This interpretation has at present no corroborative evidence to support it. The cataract and the filling in Willow Springs Draw still constitute a serious difficulty in the writer's interpretation of channeled scab-

land. The reader may consider that their existence disputes and perhaps denies such interpretation. This cannot be done, however, without a satisfactory alternative for the great gravel mounds and for the type of record shortly to be described for Yakima Valley.

COLUMBIA VALLEY BETWEEN FRENCHMAN HILLS AND SADDLE MOUNTAIN

Just below the southernmost of the three abandoned cataracts, Columbia River crosses the east-west Frenchman Hills anticline, a fold which lifts the surface east of the river an average of 400 feet and a maximum of 700 feet. Twelve miles farther south the river crosses another anticline, Saddle Mountain, parallel with Frenchman Hills. Its summit averages 1,000 feet above the broad sag between the two folds. An unnamed minor fold, about midway between these, lies in parallelism and therefore determines two minor structural valleys in the broader one. These structures, transected by the river, postdate its course. They strikingly controlled the course of glacial drainage across this part of the plateau and therefore predate the inauguration of glacial discharge. If this is correct, the Columbia Valley, throughout this stretch, and the structural valleys, where favorably situated, should also bear the characteristic erosional and depositional record. Examination has been made only near the two highways in this region, and the data therefore are fragmentary. The reader may judge whether or not they sustain the thesis of this paper.

Scabland exists along this stretch. There are no such favorable situations for its development as the broad ledge in front of the cataracts, and upper limits are difficult to determine on cliffs. The most prominent rock benches are in the transection of Frenchman Hills anticline. On the west side of the valley, scabland extends up to about 1,250 feet A.T. and has a youthful cliff back of it, like the scarps above scabland already described. The basalt flows here dip southward. If the cliff were the product of normal weathering and erosion on flows of differing resistance, it would follow the dip. Instead, this cliff retreats somewhat from the river and cuts the weaker flows above the bench at progressively higher horizons while maintaining its altitude. The scabland on the bench here is 750

feet above the river, yet very little subdued by the processes which are breaking down the buttes and filling the basins. The bench east of the river is about 1,200 feet A.T., very close to the upper limit apparently, since a pronounced and unequivocal development of channels, basins, and buttes is lacking.

At Vantage Bridge, 3 miles south of the transection of Frenchman Hills fold, no good bench exists and scabland cannot be positively identified above 1,150 feet on the cliffs.

The southern of the two minor structural valleys (now containing lower Crab Creek) carried a great glacial stream from the plateau to the Columbia, but, unlike the three spillways directly from Quincy Basin, it had no waterfall. Scabland features in this syncline at the base of the north-facing Saddle Mountain scarp have been described in an earlier paper. It will suffice here to note that the upper limit, determined about 12 miles back from the Columbia, is 1,100 feet A.T., though the preglacial structural valley there was at least 200 feet and probably 400 feet deep at that place. Glacial water was up nearly to 1,100 feet in the Columbia Valley to produce this. A similar record of the upper limit is found in the master valley for many miles downstream, yet there is indisputable evidence of great current action in many places up to this level. A current producing scabland forms of the magnitude of these, yet without apparent gradient for many miles downstream from the mouth of Crab Creek, is truly an extraordinary conception. Does not the idea literally condemn itself at the outset? But if the reader can be persuaded to read to the end of this paper, he will find that the field evidence at the mouth of Yakima Valley is as extraordinary as the conception, and that it will not permit any alternative of the flood hypothesis yet proposed.

There is a better record of deposition than of erosion by the flood water in Columbia Valley between the two anticlinal crossings. The North Central Highway from Ellensburg to Spokane descends into Columbia Valley from the west along Whisky Dick Canyon and crosses to the east side of the valley on Vantage Bridge. Coarse angular rubbly bar material lies as high as 1,000 feet A.T., 500 feet above the river, in the re-entrant containing the tributary canyon. It is mostly slope rubble collected locally, very little sorted, unworn,

poorly stratified, and very rudely foreset-bedded in places. The foresets have various dips. Here and there the deposit contains clean basaltic sand lenses from 1 to 3 feet thick. Non-basaltic fragments are rare but present. They are most abundant near the upper limit of the deposit. Much of the basalt is decayed, but there is also fresh material intimately intermingled in all exposures. The

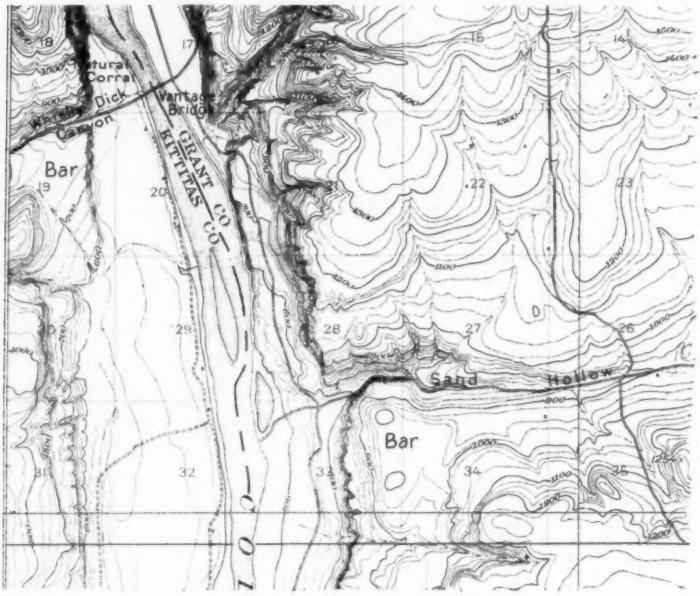

FIG. 5.—Columbia Valley just south of Frenchman Hills. From Quincy Quadrangle, United States Geological Survey.

deposit of little worn local material constitutes a complete barrier up to 700 or 750 feet, diverting the intermittent drainage off to both sides of the fill. Its riverward front is a cliff 175 feet high, cut subsequently by the Columbia. In form it suggests a huge fan built in the mouth of the tributary (see Fig. 5). A fan of these proportions, however, requires conditions that would be recorded in many, if not all, other tributaries of similar size and character in this stretch. No other tributary valley mouth deposits suggest fans.

Furthermore, the structure is quite wrong for a fan; the deposit contains material foreign to these basaltic slopes; and it extends down the Columbia along the face of the basalt cliffs far beyond the limits to which a fan from this valley could possibly grow. The original slopes of the deposit still exist in large part and are well shown on the Beverly sheet of the United States Geological Survey.

East of Vantage Bridge, the basalt cliffs are scored by many deeply gashed ravines or gulches. Virtually all of them contain irregularly disposed large mounds of angular bouldery and cobbly basalt débris, several once completely blocking the gulches. Their location prevailingly on the north sides of the gulches suggests that they are not so much a result of eddying back from the swollen main stream as of current sweeping across the spurs between gulches at, and above, the level of the deposits. One such deposit directly east of the bridge is 380 feet from base to summit. Its vertical range is not its thickness in any one section, for it lies throughout on the steep northern wall of the small coulee containing it. Original depositional slopes of these minor bars have been but little modified by subsequent drainage down the steep gulches. Foreign rock fragments occur in these deposits and scattered over the scabland surfaces up to about 1,150 feet A.T.

Sand Hollow, a small stream-eroded valley in the bottom of the northern minor structural valley, enters Columbia Valley about $2\frac{1}{2}$ miles south of Vantage Bridge. A large bar occupies the flaring mouth, constricting it below 1,000 feet A.T. to half the valley width above that altitude (Fig. 5). The bar summit is 200 feet above the valley bottom immediately upstream. Indeed, one must travel half the total length of this drainage way before the rising gulch floor is as high as this bar summit. The summit, moreover, is essentially in the middle of the flaring mouth. Later drainage has cut a narrow defile around or across the northern end of the deposit. The narrow cut descends 100 feet in about 2,000 feet horizontal, while above the bar the gradient is 35 feet a mile, about one-fourth as steep as in the defile. The bar summit includes about 200 acres and, somewhat lower, extends down the Columbia a mile beyond the mouth of Sand Hollow. Glacial water was up on the adjacent slopes at least 100 feet higher than the summit of the bar.

COLUMBIA VALLEY NEAR THE MOUTH OF YAKIMA RIVER

The great bar deposits in Columbia Valley for many miles south of Saddle Mountain gap have been described in earlier papers. Only one feature of this part of the master valley will be described here. It occurs on the west slope of the valley close to the northern base of the Rattlesnake Hills. The hills are the crest of a fold in the

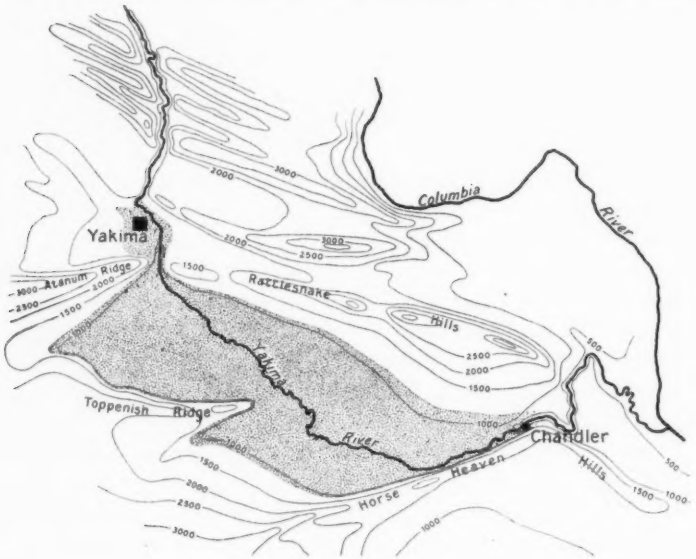

FIG. 6.—Lower Yakima Valley

Columbia basalt which trends northwest-southeast. The summits are as much as 3,000 feet above the river. Traced southeastward, the altitude decreases and the fold approaches close to the Columbia at Richland and Kennewick. Yakima River crosses a low place in the fold about 10 miles upstream from Richland. A dissected bench lies along the northeastern side of the Rattlesnake Hills below the steep descent from the crest. It abuts against the basalt fold at an altitude of about 1,500 feet A.T., from 1,000 to 1,100 feet above the Columbia. The bench is 10 miles long and from 2 to 4 miles wide (see Fig. 7). It resembles a piedmont waste slope as much as a

terrace. Its range in altitude in any cross-section is from 750 to 800 feet. No outcrops were found to show its true character, but with this we are not at present concerned.

FIG. 7.—Vicinity of the mouth of Yakima Valley. Parts of Prosser and Pasco Quadrangles, United States Geological Survey.

Over the lower part of this sloping bench occur mounds of glacial débris from 10 to 20 feet high and 100 feet or so in diameter. A dozen or more can be seen from almost any viewpoint on the northern half of the sage-covered bench. There must be several hundred of them on the slope. They are distinct constructional forms and, in places, suggest a morainic topography, though there are no associated undrained depressions, no ridge forms, and no cover of drift on the tracts separating them.

These mounds are largest and most abundant below about 850 feet A.T. The slope above carries scattered foreign boulders up to 1,100 feet A.T. Above that altitude, glacial material is wholly lacking.

The débris in the mounds is essentially a till. Striated and smoothed surfaces are common on the fragments of fine-grained rocks. The striae are mostly short and somewhat out of parallelism. Not rarely they have slight curved or angular changes in direction.

The boulders and cobbles exhibit a highly varied assemblage of rock material. Granite is the most common constituent. Slate, argillite, and fine-grained quartzite are common also. The textures, colors, thinness of stratification, fineness of grain, and presence of mud cracks and ripple marks in these fragments suggest very strongly that they came from the glaciated mountains of Beltian rock in northern Idaho, northwestern Montana, and adjacent British Columbia. Another fairly common constituent is a granodiorite porphyry, identical in the hand specimen with berg-borne boulders of this rock in virtually every scabland channel on the plateau and back up several tributary valleys along the east side of the scabland.

Very little gneiss was found, though in British Columbia, directly north of this part of the scabland, gneiss is a very common bedrock.

Five of the six mounds examined in detail contained no Columbia basalt, though a trip of at least 75 miles and perhaps 250 miles across basalt country was necessary to bring this material here, and though Rattlesnake Hills, wholly of basalt, are only 3 miles distant and rise 3,000 feet above the mound-marked area. Derivation from ice on any part of the plateau seems debarred. So does mingling of local rock waste subsequent to deposition. Yet in one mound Columbia River basalt was very common. If it did not come from the waste

of Rattlesnake Hills, it appears that some ice which melted here did contain Columbia basalt.

It seems obvious that these scattered mounds of till are great berg deposits. It seems probable that the bergs came across the scabland from the northeast. If so, they traveled at least 250 miles to reach this place. The abundance of the mounds indicates large numbers of bergs. The bulkiness of the mounds seems to bespeak bergs of enormous size.[1] If they were river-borne, great streams were required to transport them. Great abandoned glacial waterways from the northeast exist. Are they channels or valleys? If channels, were they ever filled with the enormous streams the writer conceives of? Is it only coincidence that the upper limit of this berg-borne material here is the same as that of the stream-eroded scabland in lower Crab Creek Valley, lower Snake River Valley, and Washtucna Coulee? These striking mounds on the northeast flank of Rattlesnake Hills cannot be dismissed from the problem of channeled scabland.

If these deposits are correctly interpreted as berg-laid, they must not be limited to this sloping bench in Columbia Valley. The interpretation here advanced is supported by isolated patches and mounds of berg-deposited till in many places throughout the plateau scablands and even in Columbia Valley beyond the scablands. Erratic boulders, which are more common, must mean the same thing.

It is significant that the United States Bureau of Soils map of Benton County, Washington, shows an abrupt change from one soil type to another on this bench along the 1,100-foot contour line. The subsoil profiles, as described, are notably different; that above 1,100 feet being unstratified and essentially pebbleless, while that below this altitude is stratified and with foreign pebbles, cobbles, and boulders common. The examination necessary for such a modern soil map is far more detailed than was the writer's reconnaissance. The map shows the same contact at the same altitude in Yakima Valley south of Rattlesnake Hills and, at a somewhat lower

[1] If the larger mounds are as thick as they are high, 5,000 tons is not too great an estimate of the weight of débris in one of them. If its true weight is only one-tenth of this, a berg 50×50×50 feet in dimensions would be required to float it.

altitude, 75 miles farther down the Columbia. The berg-floated débris of south-central Washington does not occur above 1,100 feet in the extensive low area where Yakima, Snake, and Walla Walla rivers enter the Columbia.

This item is essential to the appraisal of an alternative hypothesis. Berg transportation does not require river currents. Bergs may drift widely in standing water. Such a submergence has been described for this region[1] with the statement that its upper limit was about 1,250 feet A.T. That altitude is attained farther north in Quincy Basin and the Columbia Valley west of it, but the region about the mouths of the Snake and Yakima seems to show, on restudy, that no berg-floated material occurs above about 1,100 feet. Consistently throughout the scabland region, the upper limit of erratic boulders is higher farther up the valleys. Yet this increasing altitude cannot be due to subsequent warping for three reasons. (1) A theoretical "unwarping" to bring all upper-limit erratics to a common level would almost completely remove all gradient from the valleys which contain the floated boulders, and which therefore are older than the berg-drifting. (2) Such warping to explain the rising gradient of upper limit must be regularly distributed along many rather diversely oriented glacial drainage ways. No causal relations for this seem thinkable, and coincidence seems a very far-fetched idea. (3) Tributary valleys which received backwater from the scabland do not show any upstream rise in upper limit of erratics. Warping, therefore, affected only the tracts used by glacial waters. This explanation is far from satisfactory.

If the field facts justify a conclusion, it is that the upper-limit gradient of berg deposits is associated genetically, and not by coincidence, with the corresponding upper limit of loessial scarps, scabland, and bar gravel. Nowhere in the region have these berg deposits been found too high to belong to the group.

YAKIMA VALLEY

There is no region more favorably situated than lower Yakima Valley (Fig. 6) for a crucial test of the flood hypothesis. It enters

[1] J Harlen Bretz, "The Late Pleistocene Submergence in the Columbia Valley of Oregon and Washington," *Jour. Geol.*, Vol. XXVII (1919), pp. 489–506.

the Columbia more than 600 feet below the upper limit of the berg-drift débris. For 70 miles above the mouth, it is a capacious structural valley, the folding of which antedates the highest records of the flood in the adjacent Columbia Valley. The point must not be missed that its floor has been aggraded, rather than degraded, by Yakima River since the valley was made and therefore that no high-level features in it can be ascribed to early stages in valley deepening. Unlike Columbia Valley above Moses Coulee mouth, lower Yakima Valley has not been glaciated. If the Spokane Flood occurred, this valley must have received a very marked back-rush into it from the main route of flood discharge. An unequivocal record of such a back-rush is a vital requirement of the hypothesis. Yakima Valley is especially significant as a test because it was not studied at any time during the development of the hypothesis, and therefore no part of that scheme was ever constructed to take care of whatever features it might contain.

As important as the structural origin, capacity, and low altitude of Yakima Valley is the fact that it is structurally constricted at its mouth by the convergence of the Rattlesnake upwarp on the north and the Horse Heaven upwarp on the south. At the significant altitude of 1,100 feet A.T., the valley narrows from an average of 15 miles to only 3 miles, at the station of Chandler.

Not alone does this narrowing constrict the valley cross-section, but a minor fold on the southern slope of the Rattlesnake Hills, shown in Figure 7, averaging perhaps 100 feet high, actually crosses over to the Horse Heaven flexure *through the Chandler Narrows*. Thus, in addition to the narrowing, the floor of the synclinal valley is raised at this place.

In further preparation for an understanding of Yakima Valley's record, the reader should note that the lowest place in the rim of the structural valley, by which glacial water backing up from the Columbia might conceivably have escaped is 1,476 feet, 376 feet higher than the 1,100-foot upper limit repeatedly referred to. Without an escape, the flooding back must come to a standstill by the time the valley is filled. Any up-valley current which might be recorded in Chandler Narrows could have existed *only while the valley was being filled*.

Except for the flood plain and low terraces, the floor and lower slopes of Yakima Valley are buried beneath a heavy deposit of pebbly, gritty silt, generally poorly stratified and sorted, and in places without stratification or assortment. It is identical in all respects with the deposit mantling the lower part of the bench with the berg mounds (the soil map shows it as the same soil type), mantling the slopes of Walla Walla Valley to the same limits, and

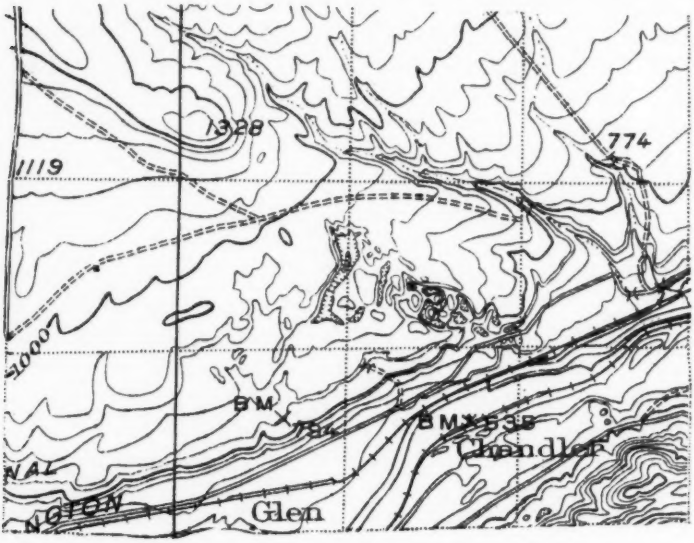

FIG. 8.—Chandler Narrows within the 1,100-foot contour

except, as to altitude, lying on the slopes of Snake River Valley and many minor valleys upstream from the scablands. Samples taken from many places show the same amazing proportions of all grade sizes from silt to fragments too large to collect, boulders in size. Boulders of Beltian rock and of the granodiorite porphyry are here. Bruises and glacial markings are common; subangular shapes are the rule.

Yet this anomalous sedimentary mantle does not go above about 1,100 feet A.T. either on the valley slopes or upstream along the valley length. Another structural enlargement, 30 miles north of Yakima (the Ellensburg part of the valley, 1,500 feet A.T. and high-

er) is totally lacking in the glacial deposit. This fact rules out the obvious alternative that valley glaciers once occupying the Yakima headwaters contributed the glacial silt. It came into lower Yakima Valley from the Columbia; it was transported upvalley.

The details of the Chandler Narrows must now be examined. Figure 8 shows the minor fold across the structural narrows, its

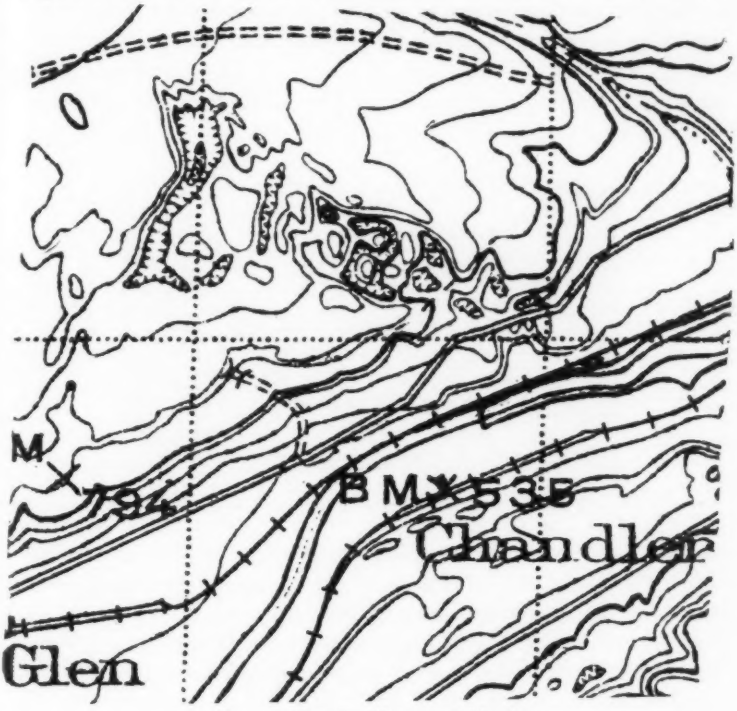

FIG. 9.—Chandler Narrows scabland

crest descending below 1,100 feet 1 mile north of the river. Closely paralleling it on the northeast is a normal erosional valley, consequent on the major structural slope. The crest of the minor fold below 1,100 feet bears a curious topography, wholly unlike anything else on the fold or on any other surfaces of the district. A closer view, magnified from the Prosser topographic map of the United States Geological Survey (Fig. 9), shows ten irregular undrained

depressions irregularly distributed among a number of hillocks. One of the depressions is more than 100 feet deep (the two inner lines, though unhachured, are depression contours) and immediately adjacent to it is a hill 100 feet high, giving a total relief of more than 200 feet in about 500 feet horizontal. Another cliff, more than 150 feet high, rises from the bottom of the largest depression.

This curious irregular topography ranges along the crest of the fold from 950 down to 600 feet above sea-level, from 450 feet above the river to only 100 feet above. It looks like a group of sand dunes or kame piles enclosing associated depressions or a cluster of sink holes. But it is none of these.

FIG. 10.—Rock basin 100 feet deep in Chandler Narrows scabland

The map alone will not solve our problem. Field examination yields the important information that the material fashioned into these sharply expressed holes and hillocks is basalt. The holes are rock basins; the hills are ragged eminences of tilted basalt flows. The problematical topography is an erosional topography in the crest of the fold below 950 feet A.T. (see Figs. 10 and 11).

Figure 12, a cross-section of Yakima Valley at Chandler, shows six different profiles: (1) the irregular topography of the crest of the fold with five rock basins in black; (2) a restoration of the original surface of the fold, drawn along its length; (3) the general structural slope alongside the minor fold; (4) the profile of the consequent tributary valley parallel with the fold; (5) the erosional valley of Yakima River in the bottom of the larger structure; and (6) the

upper limit of the glacially derived material in Yakima Valley. The floor of the deepest basin (more than 100 feet deep) is 200 feet below the crag on one side, 250 feet (or nearly so) below the crag on the other side, and only 125 feet above the river. Excavation for the highest rock basin started 450 feet above the present river level and 275 feet above the structural valley floor. All basins lie in gashes cut essentially across the fold. One is cut as deep as the adjacent tributary valley and another one is almost as deep, though both are on the structural divide.

These features are scabland features, as great a departure from normal erosional features as anything on the plateau. Scabland of

FIG. 11.—Dipping flows of the minor fold in Chandler Narrows scabland

lesser magnitude goes somewhat higher on this slope, but not quite to 1,100 feet. Chandler Narrows contains all the scabland there is in Yakima Valley, and it is significantly limited to the crest of the low fold which stood in the Narrows. Yakima River began trenching 275 feet below the uppermost of these rock basins and avoided the fold in all of its work. It cannot be held responsible. Local drainage has avoided the fold, cutting its valleys alongside, not across. One may well ask, "Is this truly erosional topography? May it not be due to collapse of lava caverns? May it not be the result of explosive vulcanism? If it is erosional, was running water the agent?"

The answer can be given from other data. The minor drainage ways down the southern slope of Rattlesnake Hills, from Corral

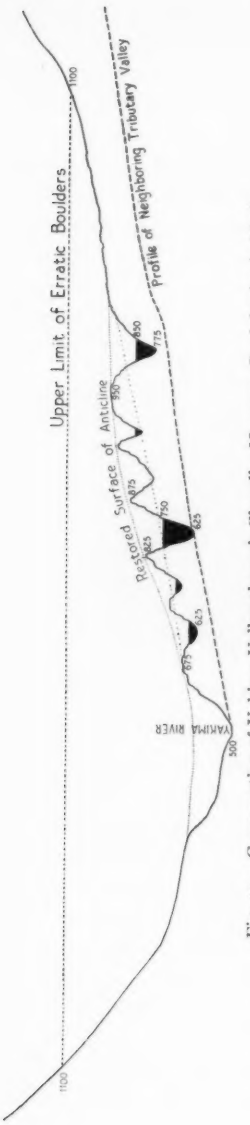

Fig. 12.—Cross-section of Yakima Valley through Chandler Narrows. Rock basins in black

Canyon on the east to Snipes Creek on the west (see Fig. 7), contain heavy deposits of little-worn gravel on their eastern sides, burying the rock walls and narrowing the valleys. The western walls lack the gravel deposits. In Corral Canyon, $3\frac{1}{2}$ miles east of the rock basins and buttes, the gravel deposit extends for 1 mile along the east wall and ranges from 700 to 800 feet A.T. In Snipes Creek Valley, $4\frac{1}{2}$ miles west of the scabland tract, this gravel buries at least $\frac{1}{2}$ mile of the eastern wall and ranges through 175 feet vertically, reaching close to 1,000 feet A.T.

These gravel deposits, well exposed in excavations, are extraordinarily bouldery in places. The boulders, cobbles, and pebbles are dominantly of basalt, very little worn to wholly angular, clearly shifted but little from the source and possessing an intimate mixture of weathered and fresh basalt throughout the total thickness. The weathered material was already altered before deposition occurred. The source yielded boulders of weathered basalt. Talus might have such a composition; so might glacial drift; but an agent which rolls its débris would soon destroy such fragments or their weathered exteriors. Sparingly distributed through the coarse basaltic waste are unweathered fragments of granite and well-rolled quartzite pebbles, derived locally from a gravelly member of the Ellensburg formation.

This gravel is foreset-bedded through-

out in every exposure. The foresets without exception dip *up* Yakima Valley, *away from* Columbia Valley to which the Yakima is tributary. The action of flowing water is recorded, but its direction of flow was exactly the reverse of normal. The foresets (Fig. 13) are not long deltaic beds, but very irregularly disposed and interrupted and cut off by overlying foresets.

More gravel deposits exist among the scabland knobs and basins of Chandler Narrows. Here they occur as separate moundings on

FIG. 13.—Foreset-bedded gravel on east wall of Corral Canyon, Yakima Valley, looking north.

the lee (upvalley!) side of buttes, in the rock basins and blocking subsequent drainage on the slopes of the group. One such bar stands 40 feet high in the bottom of the deepest rock basin and is composed largely of boulders which range up to 3 feet in diameter. Some bar forms stream upgrade westward out of rock basins and even beyond the eroded fold. There can be no reasonable doubt that the scabland and the gravel deposits of Chandler Narrows and vicinity are genetically associated. Little doubt need exist that the anomalous pebbly, sandy silt farther up the valley is also genetically associated. That the berg deposits up to 1,100 feet, both on the Rattlesnake Hills bench and throughout Yakima Valley, are a product of the same episode seems a perfectly fair conclusion. All of these features

point to a source outside of Yakima Valley and a reverse current moving up the valley.

Lest one still hesitate to admit that these extraordinary phenomena demand the extraordinary origin imputed by the writer, let us examine another narrows in Yakima Valley; one which should have the same kind of a record if the glacial water and débris came down the Yakima Valley, and which should *not* have them if the writer's hypothesis is correct. This place is the Yakima Canyon between Yakima city and Ellensburg. Here the river has cut from 1,000 to 1,500 feet deep across several folds in the basalt to escape from the

FIG. 14.—Yakima Canyon between Ellensburg and Yakima (from *Professional Paper 19*, U. S. Geological Survey).

Ellensburg structural valley and to enter the lower Yakima structural valley. The canyon is shown in the northwest corner of Figure 6, and a portion of it is shown in Figure 14. The narrowness is more pronounced than at Chandler, and this, with the winding course, should offer a much better opportunity for tremendous scour than does the Chandler constriction. But both the map and the sketch show no scabland whatever. All slopes are normal for a valley of this kind, steep cliffs outside the curves, virtually all slopes clothed with creeping waste, rock outcrops very limited, no gravel bars, no stream gravel at all except at the bottom, and that gravel occurring only in terraces. Much of the gravel exposed along the highway cuts possesses a reddened upper zone.

Whatever happened at Chandler Narrows never happened in Yakima Canyon. The floor is above 1,100 feet!

We have thus far reached the following conclusions:

1. The scabland on the minor fold in Chandler Narrows is a product of aqueous current along the major structural slopes.

2. The current action was subsequent to the trenching of Yakima River and its tributaries into the structural forms.

3. The current moved up Yakima Valley.

4. The current was tremendously effective at 950 feet, 450 feet above the present river.

5. No glacial water reached higher than 1,100 feet in Yakima Valley, 376 feet below the lowest place in the rim.

The capacity of Yakima Valley west of Chandler Narrows and below the 1,100-foot contour is 3.17 cubic miles. If there has been but one flooding of the valley with glacial water, that is the total quantity which, passing through Chandler Narrows, produced the features which have been described. The Columbia in maximum flood discharges 1,170,000 second-feet past The Dalles, about 150 miles down the river from the entrance of the Yakima. If such a flooded Columbia could be diverted into the mouth of Yakima Valley (by a blockade, let us suppose, in the Columbia canyon below the Yakima junction), it would fill Yakima Valley to 1,100 feet in about four and a half days. Any current so produced would be an uphill current, its velocity dependent on a surface gradient only and constantly decreasing as the cross-sectional area of the rising water in the Narrows increased. Since, however, Walla Walla Valley would also fill at the same time (it has the same records up to the same altitudes), the time for filling would be considerably longer than indicated in the foregoing, and the velocity would be correspondingly decreased.

What does the Columbia do today with this maximum flood over the same kind of rock, with a surface gradient of 20 feet to the mile? It has a velocity then of at least 15 miles an hour, perhaps 20 miles. The rate of erosion has not been determined, but that it is infinitesimal, compared with the erosion at Chandler Narrows, is obvious. Four and a half days of maximum flood past The Dalles have never made observable changes in the channel.

Only by tremendous velocity could a current from Columbia Valley produce these great erosional effects in Chandler Narrows

by a single flooding. Such a velocity, if conceivable, can exist for only a small fraction of four and a half days. The only conceivable way to get these results by water flowing 450 feet above the river along the slopes of a structural valley 2 miles wide at this level is to allow an advancing *wall of water*, like a gigantic bore, to sweep through the Narrows. Only thus can this vertical range of rock basins of similar magnitude, plucked out of the anticlinal crest, be produced.

Would anyone care to argue for more than one flooding to do this work? The field evidence does not debar that. But no sum of less vigorous floodings, even though they reach as high, will explain the rock basins and the mounded bars. Enormous velocity is a prerequisite: enormous velocity over almost the whole wetted perimeter of the Narrows. The shorter the time, the greater the results. To the writer, the features of lower Yakima Valley constitute the final, complete, and irrefutable evidence that there was a Spokane Flood.

The writer, at least normally sensitive to adverse criticism, has no desire to invite attention simply by advocating extremely novel views. Back of the repeated assertion of the verity of the Spokane Flood lies a unique assemblage of erosional forms and glacial water deposits; an assemblage which can be resolved into a genetic scheme only if time be very short, volume very large, velocity very high, and erosion chiefly by plucking of the jointed basalt. The total published data now encompass virtually all phases of the episode and exceed what one may readily assimilate by reading alone. It is time for a field conference with those who still insist that normal quantities of glacial water, operating through several Pleistocene episodes, could produce the phenomena. The writer does not deny a long and complicated history; he simply insists that at some time in this history there was a catastrophic flooding across the scabland, down the Columbia, and back into pre-existing tributary valleys.

THE INTERPRETATION OF INTRENCHED MEANDERS[1]

W. STORRS COLE
Cornell University

ABSTRACT

Coy Glen, one of the many gorges of the Finger Lakes region of Central New York, presents in its lower course a remarkable development of incised meanders. On first inspection, this meandering appears to be perfect evidence in support of the theory that such phenomena are a direct inheritance of a meandering course previous to rejuvenation. Further, it would seem that the initiating conditions are here fully preserved. On detailed examination, however, it is found that the Coy Glen meandering is of the ingrown type, with secondary swings controlled almost entirely by the joint structure of the bed rock. This article accordingly is intended to suggest that caution is necessary when interpreting the physiographic history of a region on the basis of incised meandering streams.

INTRODUCTION

"If an old winding stream is rejuvenated, the deepened channel follows the course which the stream had before rejuvenation. The result is that a new winding gorge is cut, that is, the old meanders are entrenched." These two sentences are from the most recently issued elementary college text[2] in geology. With more space at their disposal the authors, undoubtedly, would have added a variety of qualifications to their definition. However, as they stand, these sentences express quite accurately the commonly accepted interpretation of intrenched meanders.

The present study was made at the suggestion of Professor O. D. von Engeln, of Cornell University, who pointed out the occurrence described here, as an instance in which the fact of intrenching in rock by superimposition from unconsolidated materials could probably be definitely authenticated because the past physiographic history of the site could be read fully and accurately from existing evidence.

Detailed study and mapping, however, led to the conclusion that

[1] The author prefers the spelling "intrenched" over "entrenched" because "intrenched" was used by Rich in the first systematic classification of valleys showing such phenomena.

[2] R. T. Chamberlin and P. MacClintock, *Chamberlin and Salisbury's College Geology* (revised, 1927), Part I, p. 147.

altogether different circumstances from those postulated from inspection and preconceptions gained from the literature brought about the features noted.

This article accordingly has for its main thesis a demonstration that development of intrenched meanders does not necessarily indicate a stream previously meandering. Further, it is demonstrated that the idea of direct intrenching in consolidated rock by superimposition from loose materials is improbable rather than probable. Finally, it is argued on the basis of the evidence here present that rock structure is commonly the significant factor for the development of minor intrenched meander swings.

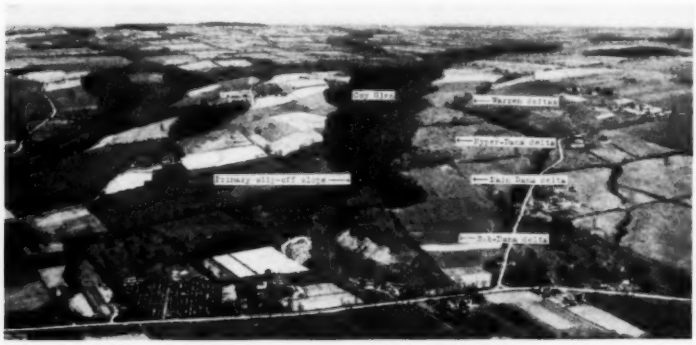

FIG. 1.—Aeroplane view to show the general relationship of the Coy Glen site. Photo by Robinson.

THE COY GLEN SITE

Coy Glen enters the main, glacially overdeepened Cayuga trough about a mile south of the city of Ithaca, New York. The Coy Glen gorge is entirely postglacial. The general relationships of the Coy Glen site are illustrated by Figures 1 and 2.

The development of Coy Glen is immediately related to the succession of proglacial[1] lakes which occupied the Cayuga Valley during the waning of the last ice of the Pleistocene invasions, and the deltas which were built in them.

[1] The term "proglacial" is used throughout the article to signify "in front of" the ice in contradistinction to "preglacial," which means "before." This term has been used by von Engeln and others in this sense.

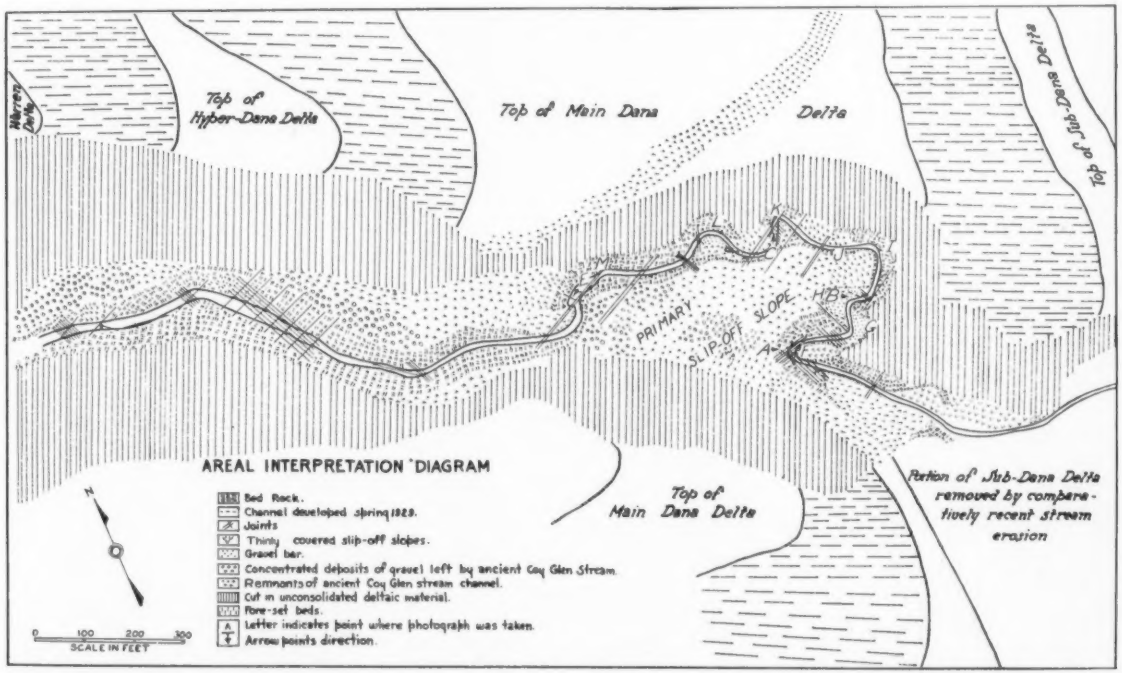
Top of Hyper-Dana Delta
Top of Main Dana Delta
Top of Sub-Dana Delta
PRIMARY SLIP-OFF SLOPE
AREAL INTERPRETATION DIAGRAM
Bed Rock.
Channel developed spring 1929.
Joints
Thinly covered slip-off slopes.
Gravel bar.
Concentrated deposits of gravel left by ancient Coy Glen Stream.
Remnants of ancient Coy Glen stream channel.
Cut in unconsolidated deltaic material.
Fore-set beds.
Letter indicates point where photograph was taken.
Arrow points direction.
0 100 200 300
SCALE IN FEET
Top of Main Dana Delta
Portion of Sub-Dana Delta removed by comparatively recent stream erosion

FIG. 2—Areal interpretation diagram

Watson[1] has identified the upper deltas at Coy Glen as deposits made in glacial lakes Ithaca, Newberry, and Warren.

He assigns three levels to Lake Warren, of which the 804-foot level is represented by the delta at the extreme west of the areal interpretation diagram (Fig. 2). The other levels represented on the diagram have not been definitely correlated by any previous worker. I consider the large middle flat extending from 615 feet to 650 feet to represent a deposit made during the main Dana stage.

The level included between 530–550 feet probably represents the sub-Dana stage of Fairchild, while that between 710–725 feet a hyper-Dana stage. The fact of a hyper-Dana stage has not appeared in the literature before except in a table published by Chadwick and Dunbar[2] in which is inserted a line: "Preliminary Lake Dana (?)."

EROSIONAL HISTORY OF COY GLEN

As the waters of the hyper-Dana stage receded, the Coy Glen stream intrenched itself in the previously accumulated deltaic material and at the same time began building the delta of the main stage. Figure 3, *A*, is a representation of the closing period of the main stage with the stream intrenched in the hyper-Dana delta flowing out over the main Dana delta. At this stage there were apparently two distributaries flowing over the delta top. The conditions on the ancient delta top probably resembled very much those of the delta which is at present forming at the mouth of Taughannock Creek in Cayuga Lake (Fig. 4).

With the opening of a new and lower outlet, the waters of the main Dana Lake in turn receded. The Coy Glen stream then began to erode the deposits made at the Dana level. The larger of the two distributaries quickly gained ascendancy, and by more rapid downward erosion cut off the water from the lesser one. The course of the lesser distributary remains as a shallow dry channel across the top of the main Dana delta. On a delta of the size of the main Dana delta, it is reasonable to assume that there was a slight initial curving of the

[1] T. L. Watson, "Some Higher Levels in the Postglacial Development of the Finger Lakes of New York State," *Fifty-first Report N.Y.S. Mus.* (1897), pp. 55–117.

[2] G. H. Chadwick and E. U. Dunbar, *Bull. Geol. Soc. Amer.*, Vol. XXXV (1924), p. 675.

stream course across its top. From this initial curving a tendency to cut laterally as well as down was given to the stream by the lower-

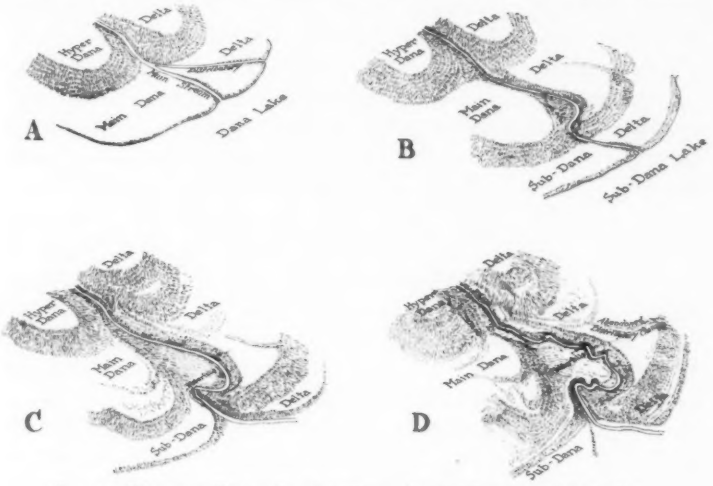

FIG. 3.—Diagrams illustrating the progressive development of Coy Glen

FIG. 4.—Aeroplane view of the delta forming in Cayuga Lake at the mouth of Taughannock (Taghanic) Creek. Photo by Robinson.

ing of its baselevel. With the initiation of the sub-Dana stage, the stream had developed a marked meander bend[1] as illustrated by the sketch (Fig. 3, *B*). Downward erosion had proceeded also, then, nearly to the level of bed rock. At the point where bed rock was first encountered there is an abrupt transition from the gentle slope of the upper western side of the valley to that of the much steeper, overdeepened, axial portion of the main Cayuga Valley. A waterfall was formed at this point which retreated upstream, forming a gorge in the rock. The sketch (Fig. 3, *C*) shows the situation there.

THE SECONDARY MEANDERS

With the initiation of erosion in bed rock, the formation of the secondary meanders was begun. Before detailed study was made, these secondary meanders were accounted for as an inheritance from the original course of the stream in the unconsolidated deltaic material or intrenched meanders, *sensu stricto*.

The primary meanders in the cover material were simply large, fairly symmetrical swings without any minor crenulation. When an analytical study of the separate secondary meanders on the site was begun, it became clear that there was an intimate correspondence between the structure of the bed rock and the position and forms of these meanders.

The secondary meanders may be divided into three classes: (1) those which are portions of the large primary meander loop and were later, after superimposition, modified by the nature of the rock structure (*F*, *I*, and *J*, Fig. 2); (2) those which owe their existence to erosional action, the pattern of which was determined by structural factors in the rock (*K*, *L*, and *M*, which is an initial meander, Fig. 2); (3) those which are developed normally from the deflection of the stream from the curved face of a meander already in existence (*G* and *H*, Fig. 2).

The term structurally controlled meanders is proposed for meanders of class 2 above. Structurally controlled meanders are defined as meanders not associated with the stream course until after superimposition on bed rock, but which develop as major or minor

[1] For a discussion of the development of initial bends into systematic curves, see W. M. Davis, *Annals Assoc. Amer. Geog.*, Vol. III (1913), pp. 5–6.

curves through the influence of the structure. The controlling structure in the instances described here is the presence of a set of master joints.

BED-ROCK STRUCTURE OF THE COY GLEN SITE

The Portage series of Devonian rocks, into which the Coy Glen gorge is being cut, is composed of thin-bedded shales and sandstones in alternation with a considerably greater thickness of shale than of sandstone. The beds have only a slight (40 feet to the mile) dip toward the south. The whole mass is strongly jointed. The principal joints direct the stream course significantly and are carefully plotted on the areal interpretation diagram (Fig. 2). These principal joints occur in two sets at nearly right angles[1] with each other. The N. 20° W. set is known as the dip joints, which are best developed in the sandy layers. The N. 65° E. are termed the strike joints, and are better developed generally in the shales.

While the N. 20° W. set is confined to the sandy layers in the vicinity of Coy Glen, the strike set cuts both the sands and shales. These strike joints occur mostly in pairs, spaced from 1 to 6 feet apart, and introduce notable lines of weakness to erosional processes acting on the rock mass.

DEVELOPMENT OF SECONDARY MEANDERS

The secondary meanders marked *F* and *I* on the areal interpretation diagram were portions of the large primary meander loop and were later modified by stream erosion and the nature of the rock structure.

The primary meander had already formed a pronounced curve at the point F. With the migration upstream of the falls referred to in the foregoing, this original curve tended progressively to take a hairpin shape because of the influence of the intersecting joint sets in guiding the course of the stream. In this section of the gorge, the rock is separated into relatively small squares, each of which is bounded by the intersecting sets of joints N. 80° E. and N. 20° W. This relation is clearly apparent by the intersection of the stream course and the joint system as shown on the areal interpretation diagram.

[1] For a complete description of the joint structure of the Ithaca region, the reader is referred to the articles by Dr. Pearl Sheldon in *Jour. Geol.*, Vol. XX (1912), pp. 53–79, 164–83.

In an analysis of the slip-off slope (Fig. 5) associated with meander at *F*, the following stages are noted. First, a moderately gentle

FIG. 5.—Slip-off slope of the meander at *F*. Note the joint planes in the bed of the stream. Photo by von Engeln.

slope composed of deltaic material, next a steep, almost vertical slope about 8 feet high marking the first bed rock and finally a very

gradual slope of thinly covered bed rock. The first moderately gentle slope represents the slip-off slope during the formation of the primary meander; the steep rock slope was formed by the migration of the first falls around this point; and, finally, the lower slope repre-

FIG. 6.—The "sluice way" falls taken from the point (*E*) looking upstream. The master joints which control the form of the falls are initiating a meander at *M*. Photo by von Engeln.

sents the migrations of the other falls upstream and a constant slip of the stream toward the cut bank. The meander at *I*, with its long, relatively gentle slope, is a modification of the primary meander curve by the undermining of the cut bank and a constant slipping of the stream toward it.

The meanders at *H* and at *G* have been caused primarily by

undercutting and lateral erosion, although the direction and the close spacing of the joints have greatly aided the stream.

The initial step in the formation of a meander such as the one at *K* is actually in process at *M*. The "sluice way" falls (Fig. 6) has its form controlled by two closely spaced N. 65° E. joints. Downstream from this falls, where these joints cut the wall, a recess is forming owing to the weathering and slumping of the rock. The stream is swinging into this hollow as it cleans out the talus, and eventually will

FIG. 7.—The meander at *K*. The master joints which initiated the meander are at the extreme left of the picture. Photo by von Engeln.

start to undermine the downstream wall. As the overhang from time to time crashes down and is removed, a marked meander will result. Figure 7 was taken looking downstream at point *K*, where the process of meander development described in the foregoing has proceeded to a marked degree. As this meander continues to grow, its general direction will always be controlled by the particular pair of joints that led to its initiation.

The meander at point *L* has developed because of the directional effect of a N. 20° W. set of joints in a sandy layer which forms the cap

rock of the small falls directly upstream from the meander site. The N. 20° W. joints here fix the form of the falls, and in consequence the water, as it leaves the base of the falls, is directed toward the north wall of the gorge.

INFLUENCE OF JOINT STRUCTURE ON STREAM COURSES

The influence of joint structure in producing a rectangular drainage pattern has been fully discussed by several authors,[1] but the development of meanders through joint control has evidently been overlooked to date. It is here argued that the phenomena described in the foregoing indicate that joint structure may as readily produce meanders as they effect a rectangular guidance of stream flow. The pattern which will result is dependent entirely on the original course of the stream with reference to the joint structure in the bed rock beneath. If the stream cuts a set of master joints at right angles or nearly right angles, meanders will result; if, however, the stream parallels or nearly parallels a set of master joints, and another set of joints almost as strong as the master joints is present at right angles to the master set, a rectangular pattern of stream course will be the result.

The directional effect, exerted by occasional joints transverse to the stream course in a layer which forms the cap rock of a falls in the course of the stream, must not be overlooked. Although a meander does not develop along such transverse joints themselves, their trend influences the main current direction below the falls, and this current impinging on the gorge wall at a point downstream starts a meander by lateral erosion. The meander at *L* is the case in point.

It will of course be obvious that such patterning of the stream flow by joint-plane influence will be effective only in streams actively downcutting, or during youth.

Where a valley is opened to late maturity of form, it has been widened by the processes of weathering as well as by the lateral erosion of the stream. The stream course has then become graded. With waterfalls eliminated, the current is not forced to take a definite

[1] W. H. Hobbs, *Jour. Geol.*, Vol. XIII (1905), pp. 363–74; P. Sheldon, *ibid.*, Vol. XX (1912), p. 79.

course; hence the process of undermining becomes incidental and variable. The valley slopes will then also be too gentle to permit recesses to be formed where the current can concentrate to round out a meander. Accordingly, joints can then play only a minor part in determining the direction of the stream's course. Hence the valley will tend to assume the open form as described by Rich.[1]

OTHER EXAMPLES OF STRUCTURALLY CONTROLLED MEANDERS IN THE ITHACA REGION

Once the phenomena of structurally controlled meanders were understood, it was quickly appreciated that less conspicuous, incised meandering courses at Filmore Glen and Lick Brook had resulted from conditions similar to those present in Coy Glen.

Possibly many of the meanders heretofore interpreted as inherited from a previous cycle of erosion will be found to owe their origin to the structure of the bed rock, regardless of the course that the stream had before superimposition.

MODIFICATIONS OF MEANDERS OF SMALL STREAMS

It is Davis'[2] opinion that by undercutting, with the resultant landslide, a stream is forced from previously developed meanders into a new course. In the case of Coy Glen, exactly opposite effects result from such undercutting because of exceptional circumstances.

At the point marked *J* on the areal interpretation diagram, the rock is some 30 feet high, and so undercut that it overhangs the stream for about one-fourth the width of the current. The sandstone layers present in the greater thickness of shale beds have proved sufficiently strong to support a wide overhang.

Accordingly, when falls of rock take place, the undermined rock mass lands on the far side of the stream, that is, on the slip-off slope. Moreover, the mass of débris commonly encroaches on the stream from that side so as to force the current to continue undercutting with even greater vigor than before. The mass of angular talus fragments now present on the lower portion of the slip-off slope indicates that such an effect has been repeatedly achieved.

[1] J. L. Rich, *Jour. Geol.*, Vol. XXII (1914), p. 469.

[2] W. M. Davis, "Meandering Valleys and Underfit Rivers," *Annals Assoc. Am. Geog.*, Vol. III (1903), p. 13.

Particular conditions of height, overhang, and rock structure are required to produce the effect described. Here these conditions are present and operate to preserve a meander instead of causing a change of course in the stream, as would normally result from such process.

Changes of the stream course and the resultant effect on meanders are effected temporarily at least in Coy Glen by trees that are

FIG. 8.—Photograph taken from the point *B* (Fig. 2). The stream has here been forced by the gravel bar (foreground) into a course which truncates the spurs, while the meander curve of the background is protected. Photo by von Engeln.

washed down during times of flood and lodge at some point in the gorge. When so lodged they act as barriers that cause gravel to accumulate to such an extent that the stream is diverted. The gravel bars shown on the areal interpretation diagram originated in this way. Because of their presence, the stream is shown at the time the diagram was made not enlarging either of the meanders (*H* and *G*) but, instead, to be truncating the noses of the spurs opposite these swings (Fig. 8).

However, during the unusually high water of the spring of 1929,

the stream opened up a course across the gravel bar as shown by the dashed line on the areal interpretation diagram. With this channel available it may be expected that the full current will very shortly be flowing in its old course again.

SUMMARY

1. With the postulated ideal conditions of a presumed earlier meandering course on unconsolidated material overlying practically horizontal beds of shales and sandstones, of which the shales form the major part, and of well-authenticated rejuvenation (here not due to uplift but to lowering of the base level), the Coy Glen stream did not incise its previously established course into the bed rock directly.

2. The large incised meanders now present in this stream are the result of lateral and downward migration, with a stream course having a very slight curvature. These primary meanders are appropriately classified as ingrown meanders.[1]

3. Secondary meanders formed after superimposition, and due to special structures in the bed rock, may be termed structurally controlled meanders.

4. In special cases, the slumping of material from the cut bank promotes, instead of reversing or preventing, meander development.

ACKNOWLEDGMENTS

The author gratefully acknowledges his indebtedness to Professor O. D. von Engeln for numerous helpful suggestions and constructive criticisms during the progress of the work and especially in the preparation of the manuscript.

[1] J. L. Rich, *Jour. Geol.*, Vol. XXII (1914), p. 470. For a complete summary of the history of the terminology applied to incised meanders, the reader is referred to R. C. Moore, *Jour. Geol.*, Vol. XXXIV (1926), pp. 44–46.

AN ANALYSIS OF BEACH PEBBLE ABRASION AND TRANSPORTATION

ROBERT E. LANDON
Butte, Montana

ABSTRACT

On the west shore of Lake Michigan between Racine, Wisconsin, and Waukegan, Illinois, beach pebbles derived entirely from a cliff of glacial till at Racine travel southward. The pebbles in the till are angular. At a point seven miles south the pebbles are spheroidal. From this point southward the beach pebbles are progressively flatter. This difference in shape is attributed to two processes, namely, wear and selective transportation. It is proved that angular pebbles become round, and round pebbles become flat. By laboratory experiments it is shown that flat pebbles travel more readily by wave action than round pebbles because of differences in behavior due to differences in shape, and to the fact that round pebbles seek deeper water and are consequently more readily buried.

I. INTRODUCTION

On the west shore of Lake Michigan, between Racine, Wisconsin, and Waukegan, Illinois, there exists a peculiar beach condition which, it was thought, would furnish reliable data concerning the wear and transportation of beach pebbles At Racine a promontory, composed of glacial till projects eastward into the lake. From this promontory, known as Wind Point, the shore recedes both to the north and south. Off-lake winds cause a southward-moving littoral current, which moves beach material southward from Wind Point, a place of active erosion. No material, however, is moved from the north side to the south side of the Point because of deep water immediately off-shore. It is clear, therefore, that the beach pebbles south of the promontory are derived from the glacial till of which it is composed, and that any change in shape or size of the pebbles is a result of transportation. A rapid survey showed that the first change in shape from the angular pebbles of the till cliff is to the rounded or ellipsoidal shape, and that as the pebbles travel farther they assume a disk or flattened shape. These observations seemed to warrant further investigation. The present paper aims to present some of the data brought out by a field study of the situation described and of a laboratory study of conditions simulating those of the beach.

II. OUTLINE OF THE PROBLEM AND METHODS OF ATTACK

The data seem to indicate that the change in shape of pebbles is due to abrasion suffered in transit, and that the change in shape is a measure of abrasion, which is directly proportional to the distance traveled, since this is constant in direction. Provided abrasion is the only factor to be considered, a study of the pebbles at various stations along the beach should give some idea of the change in shape and size per unit distance traveled. Dr. C. K. Wentworth suggested that selective transportation of favorably shaped pebbles might also be a factor. The first phase of the problem was investigated through a measurement of the shape and size of structureless pebbles, such as those composed of granite, diorite, and other igneous rocks, occurring at various stations between Racine and Waukegan. The second phase of the problem was studied in the laboratory where variously shaped objects, such as glass plates and marbles, were subjected to transportation by waves on an artificial beach. The exact methods employed in each case will be described later.

FIG. 1.—Map showing locations of stations studied.

III. SHAPES AND WEAR OF THE BEACH PEBBLES

A. METHOD OF DETERMINING SHAPE

For the description and determination of shape C. K. Wentworth's[1] method was used. The shape of the pebble, by this method, is described by two values known as the roundness and flatness ratios. The roundness ratio is the value obtained by dividing the

[1] "The Shapes of Beach Pebbles," *U.S. Geol. Surv. Prof. Paper 131-C* (1922), pp. 75–83.

radius of the sharpest edge by the mean radius of the pebble, and the flatness ratio is the corresponding value obtained by dividing the radius of the flattest surface by the mean radius of the pebble. These are designated R_1/R and R_2/R, respectively.

For the determination of R_1 and R_2 a radius or convexity gauge was used. The true mean radius of the pebble is a difficult value to obtain accurately, but the value obtained when the sum of the three dimensions of the pebble is divided by six is a fairly accurate estimate, and is fully reliable for this work. Diameters and radii are measured in millimeters.

By the application of this method to the extreme shape types, namely, the angular solid bounded by planes only, such as the cube or tetrahedron, and the sphere, the value of the method will be illustrated. For the angular solid the roundness ratio is zero, and the flatness ratio, infinity. For a sphere the roundness and flatness ratios are both equal to one. The roundness ratio is always less than one, and the flatness ratio is always more than one. In the case of the sphere the two values are equal, and in the case of the angular solid the two ratios have their extreme values. The roundness and flatness ratios, therefore, express the shape of a solid in terms of its deviation from the shape of a sphere.

B. APPLICATION OF METHOD TO LAKE MICHIGAN PEBBLES

This method of measurement and shape determination was applied to the Lake Michigan pebbles. The following table shows the

TABLE I

ROUNDNESS AND FLATNESS RATIOS OF LAKE MICHIGAN BEACH PEBBLES; TABLE ALSO SHOWS THE MEAN DIAMETERS AND DISTANCE FROM THE SOURCE

Station	Roundness Ratio R_1/R	Flatness Ratio R_2/R	Mean Diameter in Millimeters	Distance South of Cliff, Miles
1. Till cliff	0.126	12.59	36.8	0
2. Foot of cliff	0.281	5.68	64.0	0
3. Kenosha, Wis	0.382	4.69	56.8	7
4. Winthrop Harbor	0.330	6.54	34.2	12½
5. Zion, Ill	0.270	7.30	30.6	15
6. Waukegan, Ill	0.207	12.65	30.2	20

results obtained from a study of six stations between Racine and Waukegan. The pebbles were collected from the water's edge, or just

above or below, and each group of pebbles represents the average at any particular place. Only structureless igneous pebbles were used in the computations.

An examination of the data will show a rise and fall in the roundness ratio and a corresponding inverse relationship in the flatness ratio as the distance from the source increases. These data also show a progressive decrease in average size. It might be stated here that in the case of the stations closer to the till cliff, namely, 1, 2, and 3, the average size is probably somewhat greater than the figures indicate because moderately sized, rather than the largest, pebbles were collected. In the case of the other stations pebbles of all sizes were collected.

1. *Pebbles at station 1, the till cliff.*—The pebbles found in the till cliff are typical glacial pebbles. They are angular or subangular, striated, and show little evidence of having been water-worn. As shown in the table, the roundness and flatness ratios approach the ideal extremes, namely, zero and infinity. The faceted surfaces are nearly planes, and the edges between them sharp No doubt the till contains larger pebbles and boulders than those observable at the time, as large pebbles and boulders occur at station 2, immediately at the foot of the cliff.

2. *Pebbles at station 2.*—The cobbles and pebbles at this station, although they have not traveled any distance, show the beginnings of wear. Many of them still show the glacial striations and the faceted surfaces, relict features, carried over from the previous stage. The roundness and flatness ratios change considerably, showing the loss of the sharp edges and flat sides. It is evident that the wear affects the edges and corners rather than the surfaces. Any striae present are always on the large, flatter surfaces. The wear shown by the pebbles at this station has been accomplished largely in place, but may be estimated to be equal to a mile or so of travel.

3. *Pebbles at station 3.*—At station 3 the pebbles most closely approach the form of the sphere. Here the wearing of the edges has been so much as to produce a pebble with essentially the same radius over-all. This spheroidal nature of the pebbles is shown in the table by the fact that both the roundness and flatness ratios approach the value one. Individual pebbles are nearly spherical in shape, but the average values obtained depart somewhat from the ideal. The sta-

tion may not be at the exact place on the beach where the spherical form is most closely approached, but it is somewhere in the vicinity of that place. It is more probable that a zone exists in which pebbles approach the type mentioned, but in which overlaps on either side may also be found. This accounts for the fact that the roundness ratio is no more than 0.382, and that the flatness ratio is no less than 4.69. One pebble at this station, however, had the values 0.680 and 2.26, respectively, for R_1/R and R_2/R.

4. *Pebbles at station 6.*—The southernmost station visited was at the city of Waukegan, Illinois, 20 miles from station 1. The pebbles on the beach at this place are very different from any thus far described, although the ratios for the pebbles at station 6 correspond very well with the values obtained for the pebbles of the till cliff. The Waukegan beach is characterized by flat, disk-shaped pebbles, all the angular ones and most of the rounded ones being absent. Disk-shaped pebbles, with two flat surfaces and a sharp edge between them, give roundness and flatness ratios similar to those of angular pebbles. The average size, however, is less than that at any other station.

C. INFERENCES TO BE DRAWN FROM THE FOREGOING OBSERVATIONS

From the preceding observations alone it may be stated that angular pebbles become round, and round pebbles become flat, when affected by waves on a beach. It may also be stated that pebbles become smaller as they travel from their source. Table I shows that the two steps, namely, from angular to round and from round to flat, do not take place in the same distance, the former being accomplished in a relatively short distance and the other in a correspondingly great distance. It is to be inferred that the changes are toward equilibrium, and that as the stage of equilibrium is approached, the change is less rapid. The flat pebble is evidently the closest to the shape in equilibrium with the waves, since no other shape is formed after it. The reasons underlying this change in shape will be considered more fully in connection with the laboratory study next to be described.

IV. LABORATORY STUDY IN SELECTIVE TRANSPORTATION

In the foregoing discussion of the conditions along the Lake Michigan shore, it was assumed that the difference in shape of peb-

bles found at various stations on the beach is due to wear. The factor of selective transportation was neglected. Yet selective transportation may account for the groupings outlined in the previous pages, and it was the purpose of the laboratory study to determine, if possible, the nature and importance of that factor.

A. LABORATORY SET-UP

An attempt was made to produce laboratory conditions which simulated those found on the beach. In such a set-up the factor of shape could be controlled and the factor of wear could be eliminated. Other factors, such as wave velocity and depth of water, could also be kept constant. In the following experiments the variable factor was shape, and the constants were: kind of material, wear (eliminated), velocity of waves, frequency of waves, and depth of water.

A coarse sand beach, contained in a galvanized iron tank, 9 feet long, 4 feet wide, and 15 inches deep, was constructed. A galvanized iron blade, 2 feet in length and wide enough to reach into the water to a depth of $3\frac{1}{2}$ or 4 inches, and mounted on a rotating steel shaft, was placed at one end of the tank. The water was about 5 inches deep at that place. The blade was curved so that the part of the blade leaving the water at any one time was at right angles to the surface of the water. The shaft carrying the blade was placed at an angle to the end of the tank so that the waves produced could travel from corner to corner in the tank, striking the beach at an angle. Waves were produced in this way at the rate of 26 a minute.

B. LABORATORY PROCEDURE

The laboratory procedure was as follows: The sand was first heaped into the corner of the tank, not directly in the path of the waves, so that a beach with glancing waves was produced. This was done by piling up the sand in a definite position and allowing the waves to act for ten minutes. The original position of the sand heap is shown by line 1, Figure 2, and the position after ten minutes by line 2. The objects to be tested were then placed within the square marked with an x in Figures 2, 3, and 4, and the waves applied in intervals of fifteen minutes. At the end of each interval the positions of the "pebbles" were recorded on co-ordinate paper, using the sides of the tank as reference lines. In order for the pebbles to be moved

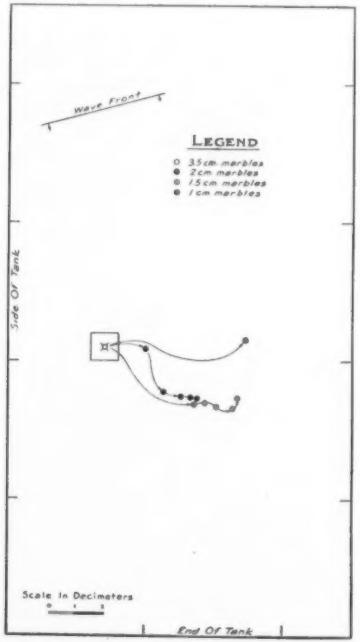

FIG. 4.—Diagram showing positions of marbles after successive fifteen-minute intervals.

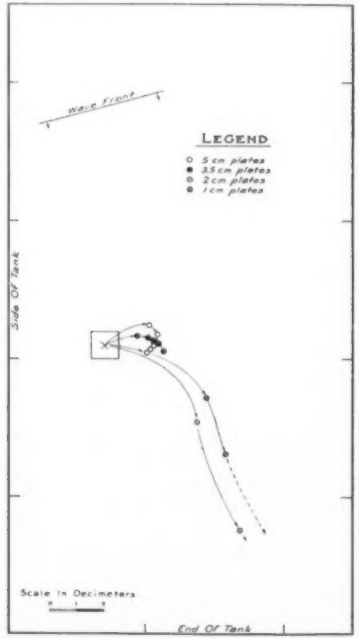

FIG. 3.—Diagram showing positions of glass plates after successive fifteen-minute intervals.

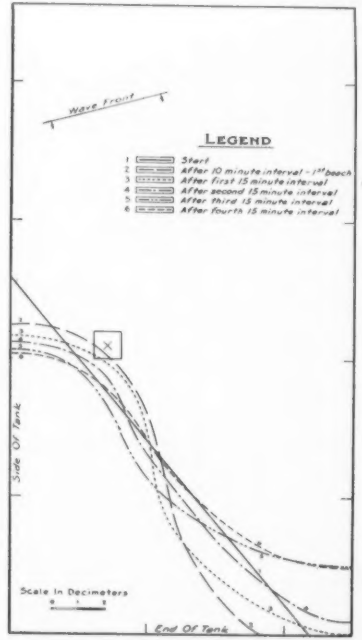
Wave Front
LEGEND
1 Start
2 After 10 minute interval - 1st beach
3 After first 15 minute interval
4 After second 15 minute interval
5 After third 15 minute interval
6 After fourth 15 minute interval
Side Of Tank
Scale In Decimeters
End Of Tank

FIG. 2.—Diagram of tank showing various positions of strand line during experiments.

the sand was also moved, and the configuration of the beach changed correspondingly. This is also shown in Figure 2.

Glass squares of the following sizes were used: 5, 3.5, 2, and 1 centimeters. Glass marbles of the following diameters were used: 3.5, 2, 1.5, and 1 centimeters. The positions of the various objects, recorded in decimeters measured along the side and end of the tank (see Figs. 3 and 4), are shown in the following table.

TABLE II

DISTRIBUTION OF GLASS PLATES AND MARBLES IN TANK DURING EXPERIMENTS IN SELECTIVE TRANSPORTATION

OBJECT	FIRST PERIOD		SECOND PERIOD		THIRD PERIOD		FOURTH PERIOD		FIFTH PERIOD	
	Side	End	Side	End	Side	End	Side	End	Side	End
5 cm. plates	11.17	5.13	10.76	5.30	10.41	5.16	10.25	5.03	*	*
3.5 cm. plates	10.79	4.61	10.61	5.16	10.59	5.30	10.48	5.27	10.21	5.67
2 cm. plates	7.7	6.9	3.8	8.4	†	†	†	†	†	†
1 cm. plates	8.6	7.2	6.5	7.9	‡	‡	‡	‡	‡	‡
3.5 cm. marbles	§	§	.....	.....	.....	.....	.....	.....	.....	.....
2 cm. marbles	10.46	5.00	8.90	5.65	8.71	6.33	8.70	6.62	8.68	6.83
1.5 cm. marbles	8.41	6.72	8.51	7.11	7.53	8.36	8.17	8.28	8.32	8.63
1 cm. marbles	10.7	8.6	||	||	||	||	.....	.....	.....	.....

* No further movement.
† Further movement not possible.
‡ No further movement because of burial in sand.
§ No movement.
|| No further movement.

In the case of the smallest glass squares the movement no doubt would have been as great as in the case of the next larger size, had they not become buried in the sand.

Aside from the horizontal or lateral separation shown in the foregoing, a vertical distribution is to be noted. The positions of the round and flat forms in relation to depth of water are shown in Figure 5. It is seen that the round forms seek deeper water than do the flat forms.

C. INTERPRETATION

The sorting process outlined in the preceding is found to be due to differences in behavior of the round and flat pebble when subjected to wave action. The differences will be brought out by a series of contrasts of the two types.

1. The flat pebble moves by "skidding"; the round pebble moves by rolling. The flat pebble stops in its motion, either with the on-coming wave or the back wash, when the volume of water is insufficient to move it. The round pebble likewise stops in its upward movement when the wave has spent its force, but in the backwash it tends to roll down the slope of the beach farther than the water itself would carry it. In this manner flat pebbles tend to stay on the beach, whereas round pebbles seek deeper water. The large round pebble in the normal wave action behaves much as a flat pebble, probably because its mass prohibits rolling by virtue of gravity alone. The large

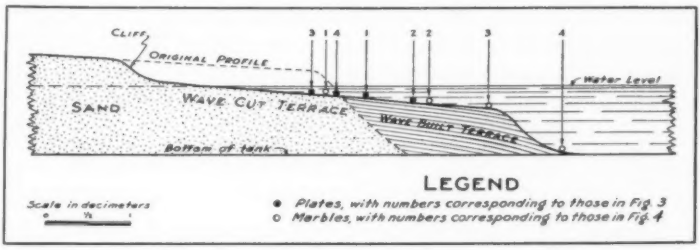

FIG. 5.—Generalized profile of laboratory beach showing vertical distribution of plates and marbles.

round pebble in times of stronger waves, however, is moved to deeper water. Small round pebbles which may be cast up by strong waves are easily returned to deep water by the normal waves following the storm. Eventually, therefore, all round pebbles find deep water, and flat pebbles are left on the beach.

2. The flat pebble is not readily lodged in gravel; the round pebble is readily lodged in gravel. The very nature of the round pebble makes it susceptible to lodgment in coarse detrital matter in deep water, but a flat pebble, on the other hand, is readily dislodged by the on-coming wave and carried onto the beach again. Round pebbles are therefore more readily buried, and temporarily prevented from traveling. Storm waves eventually bring up the round pebbles, but normal waves under most conditions have no effect.

3. Flat pebbles move more or less independently and singly; round pebbles tend to travel in groups. Since the flat pebble tends

to stay on the beach, it can travel with the normal wave, whereas the round pebble, being in deeper water, is moved only in time of storm when great masses of gravel are brought up.

4. The flat pebble is abraded differently from the round pebble. The flat pebble, traveling by "skidding," is worn mostly on its flat surfaces. The weight of the pebble is most effective in causing wear when the pebble is supported on one point, and least effective when it is distributed between two points. As a result, the greatest wear is encountered at the centers of the two flat surfaces. When the pebble is at rest the upper flat surface is exposed to the swish-swash of sand and gravel, an action tending to further flatness. The round pebble, were it perfectly spherical, would be worn uniformly over its entire surface in its rolling mode of travel, but while it is at rest and subjected to the wearing effect of other pebbles and sand, it would soon lose its sphericity. Any irregularity in the surface of a round pebble causes a concentration of wear on certain points because the rolling motion is interrupted. Any facet tends to cause the pebble to come to rest on that surface. Should it be buried, which is very probable in the deep-water zone, the last point to be covered and the first point to be exposed later, would be the point or round surface opposite the facet. Any abrasion on that point or surface tends to produce a flat pebble, which in the action of the waves after it is dislodged becomes flatter, and eventually disk shaped. It is seen, therefore, that the round pebble eventually becomes flat because of wear encountered while the pebble is at rest.

V. SUMMARY AND CONCLUSIONS

It has been demonstrated that on Lake Michigan between Wind Point, Wisconsin, and Waukegan, Illinois, pebbles are grouped as to size and shape. The largest and most irregularly shaped ones are farthest north, and the smallest and flattest are farthest south. Pebbles intermediate in size and shape are found between these two points. By field and laboratory observation it has been shown that this grouping is due partly to abrasion and partly to selective transportation, and that these two factors work harmoniously. It has also been demonstrated that wave action would eventually produce flat pebbles on any beach from any originally shaped forms.

PILLOW-LAVAS AND COLUMNAR FAN-STRUCTURES AT MURIWAI, AUCKLAND, NEW ZEALAND

J. A. BARTRUM
Auckland University College

ABSTRACT

Thick, radiately-columnar lava masses, which include large ovoid bodies as much as 80 ft. in length, are believed to represent giant "pillows" of submarine flows. Evidence suggestive of such origin is afforded by dykes nearby, which pass upwards into large spheroidal masses indistinguishable from the pillows of normal pillow-lavas.

INTRODUCTION

At Muriwai, about 19 miles due west of the city of Auckland, the bold sea-cliffs of the west coast of this northern part of New Zealand give place abruptly to a wide, sweeping, gently-shelving sand-beach, backed by a wide dune-belt which extends for mile after mile north.

A little south of the beginning of the beach, the sea-cliffs exhibit remarkable columnar structures in andesitic lavas of mid-Tertiary age, which occur sporadically in a large mass of andesitic material which builds the coastal range of hills between Muriwai and the entrance of Manukau Harbour, 16 miles south. At Muriwai itself this andesitic mass consists largely of fine-textured ejecta, including important sheets of lava, but further south it is composed almost wholly of very coarse-textured agglomerates, breccias, and conglomerates seamed by numerous thin dykes seldom more than a few feet in thickness, but including relatively few flows.

The lavas have developed pillows which are most surprisingly well displayed in the sea-cliffs near Muriwai (see Figs. 2, 3, and 4), though imperfectly elsewhere. They undoubtedly are submarine in origin, for, in addition to possessing the pillow form, in some places they include marine fossils in tuffaceous débris between the pillows, whilst they rest on tuffs with similar fossils. The seas into which they were poured, however, were very shallow, for large-scale examples of contemporaneous erosion in the subjacent tuffs are common.

In addition to pillow-like spheroids, the lavas include large radially columnar masses of somewhat unusual character, which are imperfectly illustrated by Figures 6 and 7.

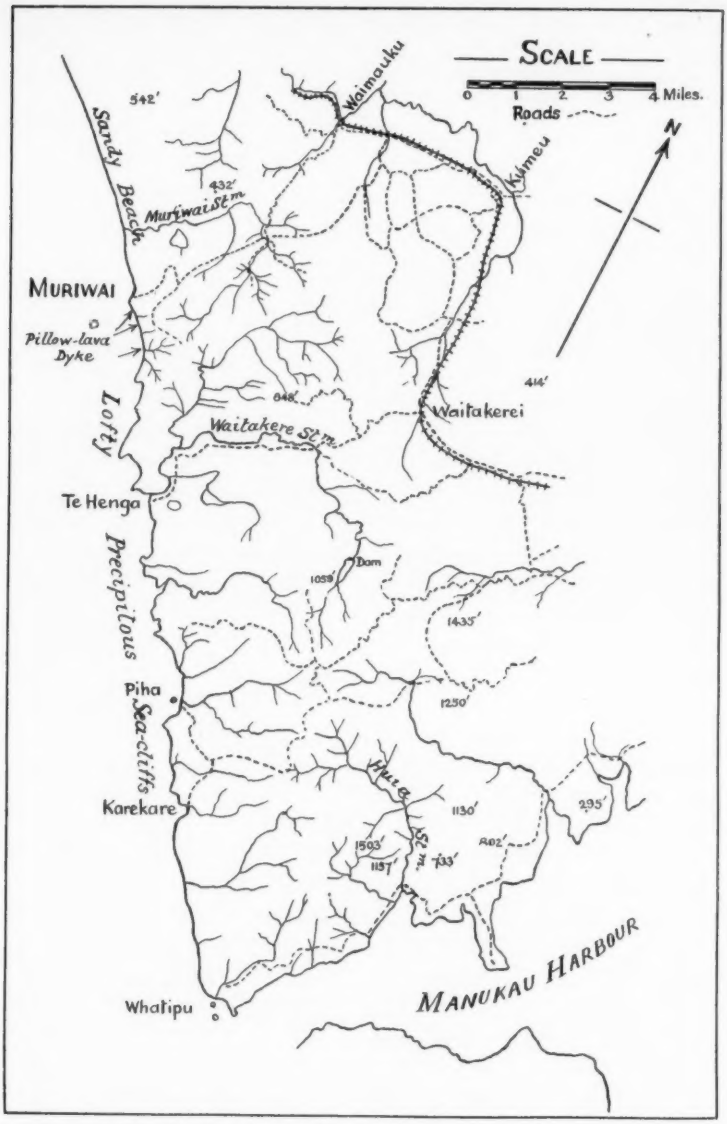

FIG. 1.—Map. Locality plan of West Coast area near Auckland, New Zealand. Muriwai is shown in the north-west part of the map.

The purpose of this paper is to proffer an explanation of the structure of the masses based on certain most striking phenomena manifested by dykes nearby.

FIG. 2.—Distant view of bluff of pillow-lava, Muriwai. Forested fore-dune at base of bluff.

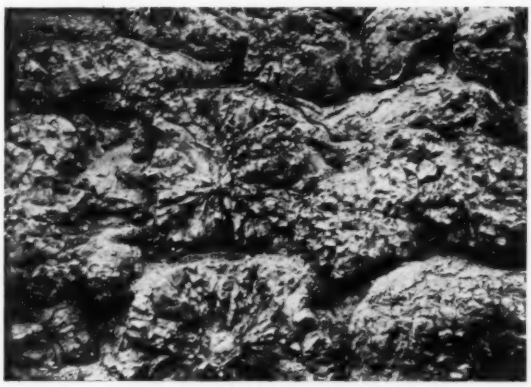

FIG. 3.—Details of pillows of main northern flow, Muriwai. Largest pillow is 4 ft. in diameter.

DESCRIPTION OF THE LAVAS

The pillow-lavas are widely exposed, interbedded with tuffs, along sea-cliffs and adjacent coastal slopes over a distance of three-quarters of a mile south from Muriwai Beach; but there are in effect, however, two main outcrops: a large lensoid northern one, with a maximum thickness of 180 ft., a little over a quarter of a mile in length; and a

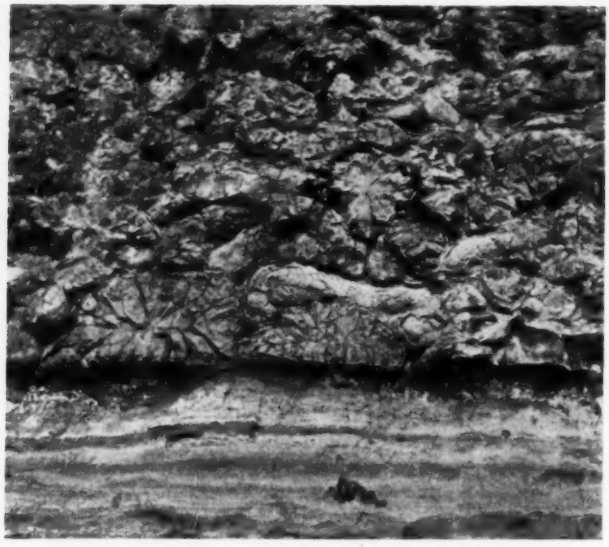

FIG. 4.—Pillows of flow of southern mass of lavas, Muriwai. The largest is 4 ft. in length.

much thinner southern one, about 14 chains in length, which is separated from the other by a gap of about 17 chains.

In the middle portion of the northern mass there is a basal flow averaging about 15 ft. in thickness, jointed in stout, sub-vertical columns (Fig. 7), which gradually thins and then disappears southwards under a mass of most characteristic pillows (Figs. 6 and 7). Above the basal flow there is a remarkable fan about 140 ft. in height and an accompanying inaccessible mass about 80 ft. in height, elliptical in section and probably ovoid in shape, which is composed

of stout columns radiating from a central core of irregularly-jointed material (Fig. 6). Inaccessible portions of the flow nearby appear similarly to be composed of columns, though disposed almost horizontally, with only the ends of the columns visible, whilst there is another gigantic radial fan, with its centre about 60 ft. above the base of the flow, exposed in the nearly vertical cliffs a little further north. Southwards, however, the columnar structures disappear and are replaced by pillows.

FIG. 5.—Dyke intruding into tuffs, Muriwai, and giving rise above to radially-columnar "pillow" about 10 ft. in diameter.

The base of the lavas preserves a surprisingly regular, gently-inclined, plane-like contact with the underlying tuffs. These latter are noticeably indurated for the depth of a few inches below the contact, whilst the lavas are chilled to a glass for about an inch above this plane, and, when the pillows are relatively poorly developed, exhibit perfect vertical tubular vesicles throughout a zone about 4 in. in thickness. These are obviously the result of evaporation of water from the basal tuffs by the heat of the covering flow rocks; they are poorly shown in the pillow-lavas.

Details of individual pillows of typical pillow-lavas are illustrated by Figures 3 and 4. Though most of the spheroids are not more than 3 or 4 ft. in diameter, others are as much as 12 ft. or more, the larger always showing good radial columnar structure, whilst all have a glassy crust about an inch in thickness. A little tuff generally appears in the interstices between pillows, especially in the lower portions of the flow.

In the southern area of their outcrop, the thickness of the lavas is reduced to an average of about 35 ft., which is very much less than

FIG. 6.—Columnar fan-structure and giant "pillow" 80 ft. high (with white centre), Muriwai. Lowest sub-horizontal columnar flow clearly visible in middle.

FIG. 7.—Columnar fan-structure and giant "pillow" 80 ft. high (with white centre), Muriwai, on left half. Below this a columnar sub-horizontal flow resting on bedded submarine tuffs.

that of the columnar portion of the extrusion of the northern area. Small perfect pillows (Fig. 4) constitute the whole of their mass, except at their extreme southern limit, where they gradually taper out as a wedge of massive material.

DESCRIPTION OF DYKES

There are two dykes which show certain extraordinary features and have almost greater interest than the lavas just described. The first outcrops 12 chains south of the main southern area of these latter, and the other about a quarter of a mile further south. The northerly one is a lensoid body, with irregular disposition, which is exposed in sea-cliffs which trend approximately parallel to its north-north-west strike. At sea-level, at its northern limit, its contact is a very steep one transverse to the strike; at higher levels it appears to occupy a curved fracture dipping gently west near sea-level, but steepening considerably as it ascends the cliff and, at the same time, gradually diminishing in thickness. Where the dyke abuts against tuffs at its northern end, it shows banded structure made prominent by weathering, whilst, as it curves upwards from sea-level, columnar jointing transverse to its walls replaces the cuboid jointing general elsewhere. Its intrusive nature is demonstrated by the details of its northern contact and by the fact that glassy, chilled phases of its rock can locally be found at its upper as well as its lower surface, with accompanying slight induration of the invaded tuffs.

At the height of about 45 ft. above sea-level, the dyke-fissure pinches out and there gives rise to a large "pillow" (Fig. 5), approximately circular in section and from 10 to 15 ft. in diameter which is composed of columns of lava radiating from its centre. This "pillow" appears to be comparable in origin with another similar spheroid developed at the outcrop of the second dyke that has been mentioned.

This latter dyke is a sub-vertical massively-jointed one which can be followed for over 60 yds. as it passes east away from the shore. Its maximum width is not obtainable, but is not less than 8 ft. at the shore, though it diminishes eastwards until, at the height of 50 ft. above sea-level, the dyke-material becomes greatly shattered and gives place immediately to a radiately-columnar lava-spheroid (Fig.

8), about 25 to 30 ft. in diameter, similar to that displayed by the dyke further north.

The lower right (southern) border of the spheroid shows a distinct, chilled selvage bordering the enclosing tuffs, but its left-hand (northern) lower and middle margin is greatly shattered where it is in contact with an irregular dike 2 ft. in thickness, from which two or three close-spaced small sills from 1 to 2 in. in depth go off for distances varying from 1 to 3 ft. This lateral dyke merges below into the shattered mass of the main dyke at the base of the spheroid, but upward it passes out of view before reaching the level of the top of the spheroid. Its rock is petrographically a basic augite andesite identical with that of the main intrusion and its pillow-like extension.

FIG. 8.—"Pillow" 25 ft. in diameter enclosed in submarine tuffs (above on left and below on right), with shattered feeding-dyke (visible on lower half) which merges with a 2 ft. dyke on left margin of "pillow."

The lava spheroid, like the earlier one described, is believed to represent the "spill over" from a dyke-fissure that reached the floor of the sea in which the tuffs of the enclosing strata were accumulating. One reason for this conclusion is that the margins of its feeding-dyke do not show glassy selvages, whilst the "pillow" has a definite, chilled crust. In addition, the material of the dyke is relatively free from vesicles, whilst that of the "pillow" has numerous coarse vesicles averaging about $\frac{1}{8}$ in. in diameter and others as large as $\frac{1}{2}$ in. in diameter.

The small lateral intrusion is regarded, therefore, as having followed the larger, earlier one after an interval that allowed the accumulation of not less than 25 ft. of tuffs.

ORIGIN OF THE STRUCTURES OF THE LAVAS

It has been shown that the phenomena associated with the two dykes described are readily explicable on the supposition that the large radially-columnar, spheroidal masses, with which they are connected, represent large pillows formed by the outpouring of small quantities of liquid rock from dyke-fissures that reached the floor of a shallow sea in which thick beds of tuff were being laid down.

If we visualise this process on a far grander scale, it is not difficult to turn from these two interesting "pillows" and their attendant feeding-dykes, and carry from them an explanation of the majestic "fans" and the 80-foot spheroid of the lavas further north, which are composed of rock indistinguishable from that of the dykes.

The spheroid is without doubt a giant pillow, whilst the masses that display the fan structures are believed more or less to parallel the domite puys of the Auvergne in their "growth by intussusception," though at Muriwai the extrusions were submarine instead of subaërial. In other words, these large columnar structures may be regarded as special aggravated examples of the pillows made familiar by ordinary pillow-lavas.

AN ALNÖITE PIPE, ITS CONTACT PHENOMENA, AND ORE DEPOSITION NEAR AVON, MISSOURI BY JOSEPH T. SINGEWALD, JR., AND CHARLES MILTON

A DISCUSSION

SYDNEY BALL
New York City
AND
JOSEPH T. SINGEWALD, JR.
The Johns Hopkins University

DISCUSSION BY SYDNEY BALL

I was deeply interested in the excellent article by my friend, Dr. Joseph T. Singewald, Jr., and his confrère, Dr. Charles Milton, in the January-February, 1930, number of the *Journal*[1] entitled "An Alnöite Pipe, Its Contact Phenomena, and Ore Deposition near Avon, Missouri." I cannot, however, agree with their conclusion under "Ore Deposition" that "there is no evidence of genetic relationship between this mineralization and the igneous intrusion," for to my mind there is strong circumstantial evidence to the contrary.

So far as I know the sulphide veinlets are confined in plan to a narrow ring around the periphery of the pipe. This alone is suggestive of a causal relationship between the veinlets and the igneous rock, although the mineral constituents of the veinlets indicate that the waters were of relatively low temperature. Broad field relationships are corroborative of this contention, for in the fluorspar district of southern Illinois and western Kentucky 120 miles to the southeast of Avon closely related and presumably contemporaneous igneous rocks are in instances closely associated with veins which produce the greater portion of American fluorspar. These veins consist of fluorspar and calcite and in instances barite with some galena and sphalerite. Certain of them indeed are rich but small zinc ore bodies. Most geologists believe the ores to be dependent upon the after-effects of the igneous intrusion.

[1] Vol. XXXVIII, pp. 54–66.

The stages in the genesis of the deposit appear to have been approximately as follows: (1) intrusion of the pipe accompanied by metamorphism of the limestone; (2) shattering of the limestone by stresses due to the cooling of the pipe; (3) deposition of ore and gangue minerals by warm waters derived from the cooling mass. As in the case of many ore deposits the evidence is not absolutely conclusive, but it is at least strong.

If such an origin be admitted for the veinlets surrounding the alnöite pipe, interesting possibilities are opened up. Slightly over a mile south of the alnöite is the Avon lead deposit which produced continuously from 1848 to 1850 and then intermittently until 1874. There are several tunnels, shafts, and open cuts which are caved or almost completely filled with wash from Saline Creek. Flat River is 15 miles west. The Avon is a typical southeast Missouri disseminated lead deposit except that the ore occurs in sandstone and in dolomitic sandstone some 50 feet below the top of the La Motte sandstone, i.e., from 70 to 100 feet below the basal beds of the Bonneterre limestone, the normal or horizon. Samples of ore on the dumps indicate that the ore was of good grade, although the ore bodies are reported to have been relatively thin. The sandstone is horizontal and the lead-bearing sandstone in many places intensely silicified. Galena with a little chalcopyrite occurs along certain bands which were evidently favorable to replacement, it being particularly abundant in the darker-colored, more carbonaceous beds. Most of the galena appears to replace the country rock. Extending up from the ore sheets are vertical stringers of galena, calcite, and some chalcopyrite. Marcasite occurs, particularly along joints in the sandstone, and this is changed by weathering to limonite. Silicification is also intense along certain vertical joints.

We are warranted, I believe, on the one hand, in considering the veinlets around the alnöite pipe and the Illinois fluorspar deposits of one and the same origin, and, on the other hand, in considering the Avon disseminated lead deposit and those of the Flat River region of one and the same origin. The data do not warrant the correlation of the geneses of the two types, although the geographic propinquity is at least suggestive.

REPLY BY JOSEPH T. SINGEWALD, JR.

The disagreement of Dr. Ball with Dr. Milton and myself regarding the genesis of the ores at the alnöite pipe near Avon arises largely from his understanding of the first sentence of the last paragraph of the section on "Ore Deposition" as a presentation of our views concerning the wider and more general problem of the origin of the disseminated lead deposits of Missouri and of the fluorspar deposits of Kentucky and Illinois. The meaning is clear only in connection with the sense of the whole paragraph.

If I understand Dr. Ball's discussion correctly he interprets that sentence to mean that we do not accept the magmatic origin of the Mississippi Valley lead, zinc, and fluorspar deposits, and the purpose of his discussion is to establish such a relationship. Reading the paragraph beyond the quoted sentence will make it clear that we are in agreement with him as to the common origin of these deposits. It will be equally clear that we did not state what our position is with respect to the magmatic or non-magmatic affiliations of that origin. This is a larger problem, not involved in the point brought out in that paragraph. I agree with him that the evidence in favor of the magmatic origin of these deposits "is not absolutely conclusive but it is at least strong." At the same time I do feel that this evidence is not as strong as for the genetic relationship of mineralization and igneous intrusion in some of the mining districts of the Western United States.

The sentence with which Dr. Ball says he is not in agreement does not then apply to the general problem of the genesis of the Mississippi Valley ores, but only to the immediate significance of the mineralization at the alnöite pipe. The three stages in that mineralization which he sets forth—igneous intrusion, shattering, and mineralization—are recognized also in our paper, with the difference that we believe the peculiar contact phenomena of this pipe indicate that the shattering was due to the violence of igneous intrusion more than it was to adjustment of subsequent cooling stresses. Dr. Ball says "the mineral constituents of the veinlets indicate that the waters were of relatively low temperature." That being the case, they might have been magmatic without an immediate relationship

to that particular igneous contact or they might have been non-magmatic. The shattered peripheral zone of the alnöite pipe would have served equally well as a favorable locus of circulation and deposition for mineralizing solutions of any origin. The relation of the mineralization to the igneous intrusion is therefore primarily structural and not causal. If the mineralization had been due to high-temperature waters, there would be good evidence also of a causal relationship between it and the intrusion. Since the waters were of relatively low temperature the causal evidence is lacking. Our statement does not deny the derivation of the mineralizing waters from an underlying magma of which the alnöite pipe is a surface manifestation. Such a broader genetic relationship, I believe, exists. The purpose of the quoted sentence and the paragraph of which it is a part was to point out that the mineralization at that igneous contact is satisfactorily explained by the structural conditions there, and that its localization there neither proves nor disproves the magmatic origin of those deposits.

Since Dr. Ball calls attention to the occurrence of the formerly worked lead deposits at Avon in the La Motte sandstone about 50 feet below the overlying Bonneterre limestone in which the disseminated lead ores usually occur, it seems worth while mentioning operations of the St. Joseph Lead Company on the old Mine La Motte property that were being carried on in 1927. East of the old mine a shaft had been sunk through the Bonneterre limestone, which was unmineralized, and extended 100 feet into the La Motte sandstone where ore was encountered. The sandstone is light gray to white in color, medium grained, porous to dense, and made up of grains of clear quartz. As it comes from the mine it is fresh looking and hard, but on the dump it soon becomes friable and disintegrates. The galena occurs in streaks and continuous bands parallel to the bedding of the sandstone. Some of it is sparsely disseminated; other bands are rich in galena. Locally chalcopyrite is encountered disseminated in the rock or in cavities with galena and dog-tooth spar. In that same year the National Lead Company was prospecting the sandstone south of Fredericktown, but I do not know with what results.

THE PRE-CAMBRIAN GREENSTONE COMPLEX OF THE JEROME QUADRANGLE. BY CARL LAUSEN

A DISCUSSION

WALDEMAR LINDGREN
Massachusetts Institute of Technology
AND
CARL LAUSEN
United Verde Copper Company, Jerome, Arizona

DISCUSSION BY WALDEMAR LINDGREN

A paper by Mr. Carl Lausen entitled "The Pre-Cambrian Greenstone Complex of the Jerome Quadrangle [Arizona]"[1] seems to call for some comment. The author has discovered "a new group of formations which is clearly younger than . . . . the Yavapai schists and apparently *younger*[2] than the Grand Canyon series."[3] This group of formations consists of volcanic flows (andesite, latite, trachyte, and rhyolite), and, as shown by the sections,[4] is beautifully folded. These formations rest on the eroded surface of the Yavapai schists as exposed near the old Shylock Mine. On the last page it is stated that "the greenstone complex is younger than the Yavapai schist and *older*[2] than the Grand Canyon series, and adds a new chapter to the pre-Cambrian history of Arizona."

There is a decided difference, Mr. Lausen states, in the physical character of the Yavapai schists and the greenstone complex. The former is recrystallized and schistose, while the basal flows of the greenstone complex show no evidence of dynamic stresses other than the folding. The author then expresses his surprise that this fact has been overlooked for so long, although a number of geologists have made examinations in this region.

Being one of this group and also noticing that my published opinions in this matter have not yet reached Mr. Lausen, I would ask the editor for permission to restate them in this *Journal*.

The significance of the unconformity so casually described seems uncertain. If it exists it may perhaps rather be as a local change of

[1] *Jour. Geol.*, Vol. XXXVIII (1930), 174–83.
[2] Correspondent's italics. Which statement is correct?
[3] *Op. cit.*, p. 175.
[4] *Ibid.*, p. 181.

facies and attitude so common in volcanic districts. I am inclined to doubt the existence of these neatly outlined folds in the greenstone complex. And I assert that the whole of the greenstone complex of the Black Hills is more or less plainly schistose, and that the schistosity is produced by the same sets of stresses applied in the same general period as those which affected the Yavapai schists. I doubt very much the efficacy of the granite intrusions to effect compressional folds and widespread schistosity. The Bradshaw granite on the whole shows little effect of dynamic stresses. Naturally the Yavapai schists, the greenstone complex, and the Bradshaw granite are earlier than the Grand Canyon series. To speak plainly, I strongly doubt whether a new chapter has been added to the pre-Cambrian history of Arizona.

In a general way the geology of the greenstone complex was quite correctly outlined by Reber[1] in a paper of great excellence.

In *Bulletin 782, United States Geological Survey*,[2] enough evidence is recorded to assure Mr. Lausen that the earlier geologists were not blind.

In the Yaeger Canyon are the greenstone series consisting of fine grained tuffs with interbedded fragmental greenstone, the series dipping 45° E. and having the usual schistosity superimposed. At the Yaeger Mine there are fissile sericitic schists, also vertical chloritic schists, and much of a massive fragmental rock which is perhaps a diabase tuff. Near the Shylock Mine the greenstones contain flat sedimentary rocks with some of the banded hematite jaspers which are so common in the Yavapai series (notes by P. C. Benedict). Where the road from Cherry Creek to Dewey crosses the schist belt from east to west the rocks consist of brown-and-red volcanic agglomerate, followed on the east (not on the west) by a wide belt of the usual chloritic schists, all members striking N. 20° E. and standing about vertical. The magnificent exposures of volcanic agglomerate in Deception Gulch and near the head of Mescal Gulch are plain enough, and at the latter place the schistosity strikes N. 20° W. while the bedding dips 45° N. Has the author considered the pos-

[1] L. E. Reber, *Trans. Amer. Inst. Min. and Met. Eng.*, Vol. LXVI (1921), pp. 3–26.

[2] Waldemar Lindgren, *Ore Deposits of the Jerome and Bradshaw Mountains Quadrangles, Arizona, Washington* (1926).

sibility of an ancient fault line extending north to south along the western foothills of the Black Hills (*op. cit.*, p. 8)?

I am afraid that Mr. Lausen does not know his "Yavapai schists" very well. His acquaintance with them seems mainly to be derived from the Bradshaw Mountain folio, by Jaggar and Palache.[1] This is an excellent and meritorious piece of work, but the fact is that the authors very seriously underestimate the amount of volcanic material (also in part intrusive) contained in the series. Is Mr. Lausen familiar with the rhyolites in the Yavapai schists in Black Canyon, at the Blue Bell Mine, at the Binghampton Mine, and at many other places? Has he seen the flat-dipping volcanic agglomerates near the Binghampton, or the amphibolites west of Mayer? It is necessary to understand that, while the variable schistosity is steep and has a northerly strike, the bedding may be very different. Commonly the schistosity is so strongly developed that the bedding is practically obliterated. Very many sericite schists develop upon close examination into pressed rhyolites.

With all these facts in view I wrote the following conclusion,[2] to which I still adhere:

> The pre-Cambrian at Jerome corresponds to the Yavapai schist of the Bradshaw Mountains, though it shows a slightly different facies. In places the schistosity is less well marked and varies in direction and the sediments are less in volume but in the main it is the same series of rocks. . . . . In general I believe these rocks [the Jerome greenstone complex] to be supracrustal—that is they originally formed lava-flows, tuffs and agglomerates. . . . . If it be admitted that the Yavapai schists in the Bradshaw Mountains also contain large masses of volcanic rocks any distinction seems futile. . . . . It is hopeless to separate them except locally.

REPLY BY CARL LAUSEN

I have read Dr. Lindgren's criticism of my article on the greenstone complex with considerable interest. The opinion of so eminent an authority is, indeed, worthy of careful consideration, and I respect his opinions. I am particularly grateful to him for calling my

[1] *Folio 126, U.S. Geol. Surv.* The repeated incorrect spelling in Mr. Lausen's paper of the name of a well-known geologist is unfortunate.

[2] *Op. cit.* pp. 20 and 56.

attention to the discrepancy in the age relations.[1] The true relations of the greenstone complex to the older and younger rocks is correctly stated in the abstract to my paper and also on the last page,[2] where the age relations are discussed.

Dr. Lindgren questions the existence of a break between the massive lava flows of the greenstone complex and the underlying schistose rocks. The collar of the shaft at the Shylock Mine is in typical Yavapai schists, and the contact with the massive flows of the greenstone complex is some distance up the hill. Here the relations are only apparent, but may be more clearly observed in a steep-walled canyon hardly more than a mile north of the Shylock Mine. In this canyon no talus covers the contact, and the thin limestone beds in brown phyllite more or less parallel the schistosity and dip to the west at an angle of 72°. The contact of the massive flows with the underlying schists is sharply defined, and dips to the east. There is no evidence of crushing along this contact to suggest possible faulting. Faulting will not eliminate the greenstone complex; it is there, and its existence must be accounted for.

I am not prepared to state how widespread is this igneous complex. When the field work for the geologic map of Arizona was being carried on, Mr. Eldred D. Wilson, of the Arizona Bureau of Mines, observed massive, dark flows of igneous rocks along the Verde River and about 35 miles southeast of Jerome.[3] He also found similar rocks along the East Verde River. He was impressed with the lack of resemblance of the flow to the typical Yavapai schists of the Bradshaw quadrangle, with which he was well acquainted, and showed them on his field maps and in his notes under a separate symbol. However, at the time this map was prepared, it was deemed advisable not to attempt to show too many separate units of the pre-Cambrian, and these black volcanic rocks were included with the schists and shown accordingly. I also examined these areas, but did not include a description of them in my paper as they were outside the Jerome quadrangle.

The exposures along the East Verde River and the base of the

[1] *Jour. Geol.*, Vol. XXXVIII (1930), p. 175.

[2] *Ibid.*, p. 183.

[3] See geologic map of Arizona, 1924.

Mazatzal Mountains are quite interesting. Ransome was the first to record a series of sediments resting on the Pinal schist. He described these rocks as follows:

> At the base of the sediments is a brick-red detrital rock made up of flakes of schist in an abundant matrix, apparently also composed of minute schist particles with an occasional grain of quartz. . . . . Overlying it without any sharp plane of demarcation is a conglomerate about 2 feet thick, with sparse pebbles of rhyolite and red jasper as much as 5 inches in diameter. The conglomerate in turn grades upward into quartzite. . . . . Little seams or films of specularite on joint planes are rather characteristic.[1]

In another traverse up the North Fork of Deer Creek Ransome again describes these sediments:

> The first rock to be seen in ascending the ravine is a large body of quartzite folded into a sharp anticline and apparently faulted against shales on the northwest. These shales resemble nothing I had previously seen in Arizona. They are gray-green, weathering yellow, fairly hard, and only moderately fissile. . . . .

Accompanied by Wilson I traced these sediments northward along the base of the Mazatzal Mountains to the East Verde River. The traverse was then continued to the west to where the sediments are overlaid by greenstones, then quartzite, and conglomerate. Beyond this point Wilson had previously continued the traverse to where the East Verde joins the Verde River. A brief account of these interesting formations is given in a report on the region.[2]

Detailed studies may disclose other areas in Arizona, and these greenstones may be quite widespread and not "a local change of facies and attitude so common in volcanic districts" as stated in Dr. Lindgren's comments.

The road from Cherry Creek to Dewey does not cross the greenstone complex as mapped by me. I fully agree with Dr. Lindgren that this road crosses a belt of chloritic schist which is a member of the typical Yavapai schist. I have seen similar rocks in Black Canyon, and between Mayer and Humboldt, which appear to have been igneous rocks originally. Similar chloritic schists occur in numerous exposures of the Pinal schist in Arizona, but it is not these rocks that

[1] F. L. Ransome, "Some Paleozoic Sections in Arizona and Their Correlation," *U.S. Geol. Surv. Prof. Paper 98* (1916), p. 158.

[2] "Gold and Copper Deposits near Payson, Arizona," *Arizona Bur. of Mines Bull. 120* (1924), p. 12.

I described as my typical greenstone complex in the Black Hills of the Jerome quadrangle.

Detailed mapping along the granodiorite-greenstone boundary shows a zone of gneisses some of which originally may have been sediments (quartz-biotite gneisses). The contact action fades out away from the intrusive mass. In the quartzites bedding planes are frequently quite distinct, and the dip is almost invariably at steep angles and away from the intrusive. It seems reasonable to assume that the turning-up of the beds of lava and the associated sediments at the contact of the granodiorite may have been due to the intrusion of magma. Other folds occurring in the greenstones may not be of similar origin, and may be due to regional stresses (p. 180).

Fortunately, beds of quartzite and argillite are quite common in places in the greenstone complex, and numerous dips were recorded. It was possible, therefore, to show the major folding. Minor crenulations may occur, but they would be difficult to show on the scale of the drawing.

The region in which these greenstones occur is quite accessible. The main highway to Jerome crosses a belt of these rocks in Yaeger Canyon. The highway from Dewey to Cherry Creek crosses a belt of typical Yavapai schist but does not cross rocks which I have described as the greenstone complex. However, branch roads to prospector's claims lead northward from this highway into the very heart of the region under discussion. The student of pre-Cambrian geology is cordially invited to visit and study this interesting region and to add his observations to those already recorded.

PRE-CAMBRIAN WATER-LAID TUFF IN THE BARABOO, WISCONSIN, DISTRICT

J. T. STARK
Northwestern University

ABSTRACT

An outcrop of pre-Cambrian rock in the Devils Lake area of south-central Wisconsin, described in the literature as a rhyolite flow, is believed to be a fragmental, water-laid tuff.

In visiting the isolated outcrops of pre-Cambrian rocks in the Devils Lake area of south-central Wisconsin in 1929, the writer was impressed with the resemblance of some of the so-called rhyolite flows to the tuffaceous slates of Huronian age in northeastern Minnesota. Further field investigation strengthened the idea that these were fragmental, water-laid deposits. Microscopic study has supported such an origin for at least one of the outcrops.

The geology of the Baraboo iron district around Devils Lake is well known for its interesting physiographic, structural, and economic history, all of which has been ably described in reports and bulletins of the United States and Wisconsin Geological surveys.[1] The present paper is believed to be a new if slight contribution to the geology of the region.

Only a brief review of the principal structural features of the pre-Cambrian formations is necessary. An east-west syncline or synclinorium is outlined by dissected ridges of the Baraboo quartzite, presumably of Middle Huronian age, overlain conformably by the Seely slate and Freedom dolomite. The lower flanks of the ridges and the surrounding plain are covered with the horizontally lying Cambrian (Potsdam) sandstone. The fold is canoe-shaped, asymmetrical, extending 25 miles east and west and ranging from 2 to

[1] R. D. Salisbury and W. W. Atwood, "The Geography of the Region about Devils Lake and the Dells of the Wisconsin," *Bull. Wis. Geol. and Nat. Hist. Surv. No. 5* (1900); Samuel Weidman, "The Baraboo Iron-bearing District of Wisconsin," *Bull. Wis. Geol. and Nat. Hist. Surv. No. 13* (1904); C. R. Van Hise and C. K. Leith, "The Geology of the Lake Superior Region," *Mon. U.S. Geol. Surv. No. 52* (1911), pp. 359–65.

12 miles in width. The dips on the north flank approach vertical and on the south flank they vary from 10 to 35 degrees.

Just outside the borders of the syncline occur the pre-Cambrian igneous rocks in isolated patches of rhyolite, diorite, and granite. Weidman has described these in some detail, and has shown convincingly that they form the floor upon which the overlying sedimentary rock, the Baraboo quartzite, was deposited.[1]

The particular outcrop discussed in this paper occurs on the Markert farm, a mile and a half northeast of Denzer, in the S.E $\frac{1}{4}$ of Sec. 11, T. 10 N., R. 5 E. Loose blocks of true rhyolite were found near the road as described by Weidman; but in the outcrops along the stream about 150 yards above the bridge, the rock appears distinctly different from the loose blocks. Rock in place is exposed for about 200 feet with a width of from 5 to 20 feet.

The rock has two distinct facies, varying from a light gray color with a faint greenish cast, of massive slatelike texture, to a black, extremely dense, hornfels-like rock with coarser bands, resembling Huronian graywackes of the Lake Superior region. The general appearance of the exposure is not unlike many outcrops of the Knife Lake slates on the Vermilion Range in northeastern Minnesota. It was this similarity that first suggested a water-laid deposit. Roughly parallel bands, which might easily be mistaken for flow lines of a rhyolite, on closer inspection give the appearance of distorted bedding. The rock is highly sheared and distorted with many small offsets, so that dip and strike directions in so limited an exposure may mean very little. The banding has a general trend of N. 38° W., and a dip to the northeast of approximately 73°. If this banding is true bedding, such a strike is nearly at right angles to the strike of the south flank of the quartzite syncline, outcrops of which occur just to the north.

In thin sections the true rhyolite of the loose blocks, none of which was found in place in this vicinity, shows the characteristic texture and composition of a recrystallized surface flow (aporhyolite). Numerous phenocrysts of feldspar and pyribole are imbedded in a finely crystalline ground mass of the same composition. The feldspars form the most abundant phenocrysts. All are highly altered to

[1] Weidman, *op. cit.*, pp. 20–21.

masses of sericite, chlorite, epidote, and carbonate. Those showing twinning were determined as oligoclase. Original pyribole phenocrysts, next in importance, are represented by completely altered areas of chlorite, epidote, and carbonate. They are probably the remains of original hornblende, but none of the primary mineral was found. Quartz phenocrysts in clear, rounded grains are scattered sparsely throughout the sections. The entire rock shows innumerable small grains of secondary sericite, chlorite, epidote, and carbonate. Crystal outlines in many of the phenocrysts are clearly recognizable. In general the rhyolite of these loose blocks, but for the absence of flow lines, resembles closely in composition and texture the rhyolite occurring in other parts of the region.

The coarser facies of the massive, slatelike rock show distinct fragmental textures under the microscope, Figure 1. The larger fragments are sharply angular and consist mainly of orthoclase and oligoclase feldspar. In general the feldspar is fresher than in the true rhyolite. Areas of chlorite and epidote suggest alteration from pyribole fragments. Crystal outlines are not evident as in the phenocrysts of the rhyolite. Some of the feldspars are completely sausseritized and sericitized. The larger fragments are imbedded in a finely crystalline, recrystallized groundmass of interlocking grains of quartz and feldspar. A few crystals of secondary apatite are present. In addition to the mineral fragments, there are rock particles which in some instances show true phenocrysts and appear to be from rhyolite similar to that of the loose blocks. Other of the rock fragments resemble fine-grained quartzite or slate. The rock is a lithic tuff.

In the dense, hornfels-like facies the rock is composed of very fine, interlocking crystals apparently of feldspar and quartz. Areas of sericite and chlorite spot the sections but do not suggest altered phenocrysts. Their outlines are extremely irregular, suggestive of similar development of such minerals in metamorphosed slates. Secondary pyrite in cubic crystals occurs in many of the sections, showing alteration borders of hematite. Parts of the rock show fine laminations, and in thin sections these are seen to be composed of regular bands of finer and coarser grains with gradations from one to the other, Figure 2. The regularity of the banding seems to be

aqueous in origin. Contacts between the finer and coarser bands are distinctly those of a sediment rather than flow lines of an igneous rock.

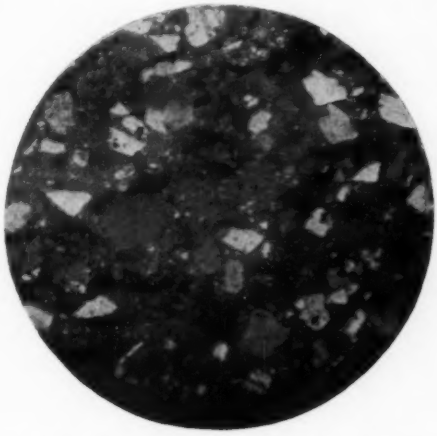

FIG. 1*a*.—Coarse facies of the massive, slatelike rock showing fragmental texture. Without analyzer, ×60.

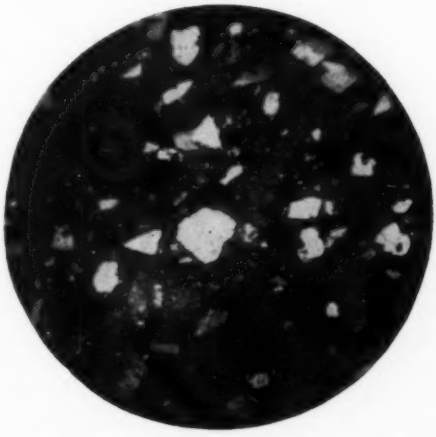

FIG. 1*b*.—Coarse facies of the massive slatelike rock showing fragmental texture. With analyzer, ×60.

In summarizing the reasons for thinking that the Markert farm outcrop is a fragmental deposit of tuffaceous origin, and partly at least laid down in water, the following differences from true rhyolites of the region may be listed: (1) the sharply angular character of the feldspar and quartz fragments; (2) fragments of rock mixed with mineral particles; (3) fine laminations, regularly parallel, and

FIG. 2—Dense, hornfels-like facies of the rock showing stratification. Without analyzer, ×30.

in no way suggestive of flow structures; (4) suggestions of cross-bedding on the weathered surfaces of the finer grained facies of the rock; (5) the character of the contacts between the finer and coarser bands, the one grading into the other as in water-sorted sediments. The coarser, massive facies of the rock, while clearly tuffaceous in their fragmental textures, do not show any evidence of sorting or deposition in water. It may be that this material was deposited so rapidly that no sorting took place, or else it may have been a true subaërial tuff. In the denser facies, however, the evidence points

unmistakably to aqueous sorting, and in view of this fact it seems probable that both facies were deposited in water.

The age of the tuffs is very probably contemporaneous with the volcanic rhyolite flows of the region, that is, definitely before the invasion of the sea in which the Baraboo quartzite was deposited. They may represent a portion of the pre-Proterozoic complex, or possibly they were deposited during Lower Huronian time.

The writer wishes to acknowledge the help of Professors W. H. Haas, H. B. Ward, and Mr. W. E. Powers, who were with him in the field on the first visit to the outcrop, and of Professor U. S. Grant, who was with the writer on a second trip to the area.

REVIEWS

Geology of the McCalls Ferry–Quarryville District, Pennsylvania. By ELEANORA BLISS KNOPF and ANNA I. JONAS. *U.S. Geological Survey Bull. 799.* Washington, 1929. Pp. 153; figs. 15; pls. 8; geologic map (in pocket).

The McCalls Ferry–Quarryville district, lying along the lower reaches of the Susquehanna River and embracing southern Lancaster and part of southern York counties, is of particular interest because it bridges a gap between the relatively legible structure of the Appalachian Paleozoics and the difficultly decipherable belt of intrusives and metamorphics lying just inland from Philadelphia and Baltimore.

Of more than local interest are the authors' correlations which are based on their extended regional studies. The oldest formation, the pre-Cambrian Baltimore gneiss, they consider approximately the equivalent of the Grenville. Their Glenarm series, they believe to be later pre-Cambrian corresponding with the Manhattan schist of New York and its underlying limestone and quartzite. The Chickies quartzite, which is early Cambrian, they correlate with the Hardystone and Poughquag quartzite of New Jersey and New York. The Cambro-Ordovician limestone series is the Shenandoah and Wappinger of the Appalachians while the Cocalico shale is equivalent to Martinsburg and Hudson River.

In this portion of Pennsylvania, one of the most puzzling anomalies has long been the remarkable behavior of a belt of slates and schists. They overlie the Cambro-Ordovician limestone of the Chester Valley and, farther south, are cut by pre-Cambrian intrusives! If, as seems highly probable, the limestone is really Cambro-Ordovician, it is obvious that either (1) there are two kinds of schists (one Ordovician and one pre-Cambrian), or (2) they do not truly overlie the Cambro-Ordovician limestone, or (3) the intrusives are not pre-Cambrian. Dr. Florence Bascom chose the first way out of the difficulty and inserted a fault between the phyllites directly overlying the limestone and the more highly metamorphosed mica-gneiss in the intruded region. On carrying this structure west into the Coatesville-Avondale district, Bliss and Jonas, her students and colleagues, found this pre-Cambrian gneiss resting, over a large area, upon a limestone south of the Chester Valley which they considered equiv-

alent to the Cambro-Ordovician. Their recourse was to make Bascom's fault do duty as a very flat overthrust subsequently folded and worn through by erosion. Then Hawkins[1] pointed out a difficulty (which had also puzzled the reviewer): The contact between the two kinds of schist (Octorara, Ordovician, and Wissahickon, pre-Cambrian) is gradational and shows no signs of faulting. Accordingly, he explained the gradual change from phyllite to mica gneiss as due to varying degrees of metamorphism. This view has been adopted by Knopf and Jonas in their latest publication, though no doubt they arrive at it independently, as the bulletin under review does not credit Hawkins with proposing it. They have therefore moved the flat overthrust plane downward, placing it between the limestone and the schists and making the whole schist belt, which they call the "Glenarm series," a great pre-Cambrian wedge, thrust northward twenty miles or so, though they candidly admit that they have been unable to find any direct evidence of faulting along the upper contact of the limestone. In fact, Bascom cited evidence, quite convincing at the time, for believing that in the Philadelphia district the schists rest conformably on the Shenandoah limestone.

One might offer in objection to the proposed fault the fact that the Glenarm series does not appear between the Baltimore gneiss and the Cambrian quartzite north of the Chester Valley, as it should if it is pre-Cambrian.

These objections the authors have foreseen and explained in a very plausible manner. One suspects, however, that their reluctance to cut the Gordian knot and make the Glenarm series all Ordovician rests on the rather revolutionary consequences that such a procedure would entail, for in that case the widely distributed gabbro and ultra-basic rocks in Pennsylvania and Maryland, as well as some of the more acid intrusives, would be Paleozoic. But this would at once explain the locally intense metamorphism of the Paleozoics as well as of the Glenarm series, provide a tangible instead of a hypothetical source for the recognized post-Cambrian pegmatites, and avoid the necessity of invoking a flat, folded overthrust. After all, just how sacred is the tradition of pre-Cambrian age for the intrusives? Is it not rather remarkable that the great period of Appalachian orogeny was accompanied by no known intrusives in this region? Not that the idea is without its difficulties, chief among which is the discovery by Miss Jonas that basal Cambrian rocks rest conformably upon the Glenarm series in Frederick County, Maryland, and York County, Pennsylvania. It is the reviewer's opinion that the difficult problem

[1] A. C. Hawkins, *Amer. Jour. Sci.* (5th ser., 1924), Vol. VII.

is not yet solved, despite the authors' masterly treatment of a large volume of field and laboratory information.

An interesting section is devoted to the physiographic record in which the work of Bascom, Barrell, and others is extended and modified. The idea of a late Cretaceous peneplain is discredited: "No erosion cycle between Blue Mountain and the Coastal plain is older than late Tertiary." While disproving the conception of numerous cycles of complete peneplanation of the area, the authors recognize no less than eight successive incomplete erosion cycles, indicating a periodicity in the gradual uplift of the area.

H. E. McKINSTRY

Mineralogy: An Introduction to the Scientific Study of Minerals. By SIR HENRY A. MIERS. Revised by H. L. BOWMAN. London: Macmillan and Co., 1929. Pp. 658; figs. 729. $8.50.

The appearance of this well-known work in revised form is of great interest. The scope of the changes made is stated in the Preface as follows:

> In the new edition the original plan of the book, and so far as possible the original text, have been retained, with such changes and additions as seemed desirable in view of advances in the subject or to make the book more useful to students.
>
> In the chapter on the optical properties of crystals the idea of the wave-surface has been introduced, as it probably enables a student to form a clearer picture of the passage of light through a crystal than he can gain from the ellipsoids alone.
>
> The section on crystal structure has been somewhat enlarged and a chapter added on X-ray analysis.
>
> A considerable number of new figures have been added, and many of the oid ones redrawn, while the whole of the text has been carefully revised.

The new edition has 74 more pages than the old, and has been reset. Outside of the changes noted above, the increased pagination is accounted for by numerous small but worthwhile additions or restatements well scattered through the book. Detailed attention cannot be called to these in a short review.

Some 500 pages of this work are split nearly equally between crystallography and descriptive mineralogy. Although the former is broadly covered, to the reviewer it seems that morphology is very heavily stressed; this is perhaps most apparent when examining the descriptions of the minerals. Thus the zeolites are accorded 12 pages, while clay minerals are covered in slightly more than a single page. Boracite takes more than 2

pages (peculiarly the very interesting crystals of colemanite have only a paragraph), whereas bauxite is given but 4 lines. Bournonite has nearly 2 pages; bornite (erubescite) a half-page; magnesite but a sentence. Glauconite is not described, though it is listed in the tables near the end of the book.

As stated in the original Preface, the object sought was "a treatise which can be *read* by the student." Probably morphology is emphasized for the same purpose; namely, to arouse the interest of the thoughtful student. From such a standpoint this well-illustrated work must be regarded as highly successful in all respects, except that a description of the fascinating Goldschmidt two-circle method of measuring, calculating, and drawing crystals is omitted from the short chapter on goniometry.

The descriptive part of the work is inadequate as regards genesis, geologic occurrence, and alteration products of minerals (although "these constitute the most important and, perhaps, the most interesting section of Mineralogy proper") because these subjects "deserve to form a separate volume." Of course, to cover these completely might require several such volumes, but the reviewer considers that the space given to "Noteworthy Localities" might be devoted to considerable advantage to a brief treatment of these subjects, allowing the teacher to cover localities, if he felt that lectures dealing with descriptive mineralogy were advisable.

Human fallibility prevents the preparation of a compendium with so much factual material that is free from errors. The present work is remarkably good in this respect. Some of the things that might be placed under this heading which were noted during a brief examination follow: The didymium given in the formula of monazite is now known to consist of two rare-earth elements, praseodymium and neodymium. As two kinds of iron are shown by the magnetite formula, so should two kinds of copper be indicated in the formulas of enargite and tetrahedrite (and bornite, too, as recently given by Gruner). Covellite is not always microcrystalline. Andalusite or cyanite when "raised to a high temperature" invert to mullite, not to sillimanite. The latter mineral also undergoes this inversion. The statement that "both nitre and nitratine are probably to a large extent of organic origin, being formed by the action of nitrifying bacteria in the soil on nitrogenous organic matter" hardly gives a fair picture of the great desert of Chile to an embryo student. It is to be regretted that some of the excellent work done at the Geophysical Laboratory in Washington on the phase relations of the silica minerals, as well as Holden's interpretations of the colors in quartz, is not adequately presented. It is stated that "the quartz from Herkimer County encloses an-

thracite"; also that quartz, tridymite, and cristobalite "are mutually transformable with ease at definite temperatures." Quartz is hardly "the most common" mineral, as one or two of the feldspars are more abundant. More complete discussions (and diagrams) of the optical properties and chemical relationships of some of the leading groups of silicate minerals might have been included to advantage. (Feldspars are so treated in the first part of the text.)

The reviewer has wandered far afield and has been quite unfair to the book supposedly being described in an effort to present briefly his conception of the ideal textbook of mineralogy. The object of the author as stated in the third paragraph above has been achieved, as all English-speaking mineralogists know. Those who teach will find a copy of this new edition a valuable addition to their private libraries.

D. J. FISHER

Die Geschichte des Vesuv: Illustriert nach gleichzeitigen Urkunden ("The History of Vesuvius: Illustrated from Contemporary Documents"). By G. B. ALFANO and I. FRIEDLAENDER. Berlin: Dietrich Reimer, 1929. Pp. 71; pls. 107; maps 2. Price M. 50.

This elaborate work is remarkable in a number of respects, and is a notable achievement. The activities of Vesuvius have been under observation for many centuries, and published descriptions run into the thousands. The present authors have been intimately connected with observations of the volcano for several decades, and the institutions with which they are associated (Pius X Observatory and Immanuel Friedlaender Volcano Institute) possess rich collections of books, paintings, engravings, and photographs. From these they have selected material for a history of Vesuvius from the earliest times to the present, using, as far as possible, works of authors who were eyewitnesses of the occurrences, and avoiding fantasy. Such selection involves much research and discrimination, but it has been well done. The sequence is historical. To each chapter a bibliography is appended.

Beginning with the Plinian eruption of 79 A.D. (or even going back to faint evidences from prehistoric times), a record is presented which includes not only the salient phenomena of the great eruptions, but also the minor activities and the state of the volcano in relatively quiet intervals. The narrative is chiefly descriptive, and very little in the way of interpretation is attempted. To most readers the major phenomena will be of greatest interest, and some of the descriptions of quiet periods appear

hardly more than a dry cataloguing of conditions, but they are essential for a complete record.

Emphasis is placed on the changes that the volcanic edifice has undergone as a result of destructive and constructive operations, and some of the changes are striking. For example, evidence is given that during the long period of perfect quiescence prior to 79 A.D. the mountain possessed only one summit, accordant with the outer slopes of the present Somma; that this was blown away in the eruption, and that the present great cone is of later construction. Subsequent changes have involved repeated breaking-down and building-up of the later summit and inner cones, increase and decrease of diameter of the throat, opening of fissures, and the formation of more or less transient fumaroles.

The volcanic manifestations, simultaneous or at different times, have been of astonishing variety: lava flows from the summit and from lateral openings, discharges of lapilli and great blocks, ash rains (dry and wet), fire fountains, gas discharges, sublimation of volatile salts, mud streams, water torrents, avalanches, "pine" clouds, great explosions, earthquakes, whirlwinds, lightning, fluctuations of sea-level, drying-up of brooks, and dislocation of the ground.

Some phenomena are so difficult to interpret, even approximately, that their very existence has been denied by some. Such is the discharge of great masses of water from the mountain during the outbreak of 1694, after the eruption was well under way and many explosions had occurred. Consideration of the phenomena in general indicates to us how little explanation, other than surmises, volcanologists have to offer as to fundamental causes of volcanism, or even of superficial manifestations.

The volume is illustrated with 107 plates, reproduced (many in colors) from oil paintings, gouache pictures, engravings, and photographs. These have evidently been carefully selected, not only with the purpose of presenting faithful and vivid representations of the spectacular occurrences, but also with an eye to artistic merit. Their reproduction (especially of those in colors) is of high excellence, and they will give pleasure to many persons not scientifically inclined. The expense of their production explains the high cost of the volume (M. 50) and the limited edition (200 copies). It is a book for libraries, reference collections, and specialists rather than for the ordinary individual.

It is accompanied by two carefully drawn folding maps, one of the mountain and its immediate vicinity, and the other of the surrounding region, which show, in great detail, lava flows, lines of dislocation, and Roman and modern cultural features.

C. N. FENNER

Horton-Windsor District, Nova Scotia. By W. A. BELL. *Geological Survey, Canada, Memoir 155.* Ottawa, 1929. Pp. 265; text figs. 14; pls. 36.

The results of the detailed study of the Horton-Windsor district, which was begun by Dr. Bell nearly twenty years ago and interrupted by the World War, are at last, after many delays in the printing, made available by the publication of *Memoir 155*. The investigation was primarily of the fauna of the Windsor series in the type area, but the *Memoir* is somewhat larger in scope as it treats also of the general geology of the area.

The very detailed work of Bell has shown that the Windsor faunal succession may be tabulated as follows:

UPPER WINDSOR

Martinia zone

Characterized by *Martinia galataea, M. thetis, Composita obligata, Lophophyllum avonensis,* and *Phillipsia eichwaldi*

c) *Caninia dawsoni, Zaphrentes minas, Chonetes politus, Tabulipora acadica, Spirifer nox*

b) *Productus semicubiculus, Composita obligata*

a) *Nodosinella priscilla, Lophophyllum avonensis, Dibunophyllum lambii, Spirifer adonis*

LOWER WINDSOR

Composita zone

Characterized by *Composita dawsoni, Diodoceras avonensis, Batostomella abrupta, B. exilis, Septopora parva, Diaphragmus tenuicostiformis,* and abundant *Productus* of the type *corrugata-hemisphericus*

The Lower Windsor thus shows a striking faunal similarity to the Middle Visean (S_2) of the Avonian sequence; and the equivalency of the Upper Windsor with the Upper Visean is indicated by the appearance in both of the characteristic genus *Dibunophyllum*. Although the European affinities of the Windsor species are marked, the relationship of the latter to those of the Mississippi basin is most casual. The Lower Windsor, however, is pretty certainly a correlative of the earlier Chester, though with some slight suggestion of Ste Genevieve relationships; and the Upper Windsor probably was deposited contemporaneously with the Upper Chester. In spite of these rather definite correlations, practically no Windsor and Chester species are conspecific—a fact which indicates clearly the degree of isolation of the two areas of deposition.

The Windsor fauna, as described by Bell, consists of 127 species belonging to 81 genera. Of these, 56 species and 2 brachiopod genera, *Protoniella* and *Harttella*, are new. The assemblage is dominantly molluscan-

brachiopodan, as is evidenced by the fact that the brachiopods are represented by 26 genera and 49 species, and the molluscs by 34 genera and 55 species. The common Chester bryozoan *Archimedes* is conspicuous by its absence, and not only is *Pentremites* unrecorded, but not a single echinoderm has been found!

The *Geology of the Horton-Windsor District* is an important contribution to the geology of the Canadian Maritime Provinces. The report is beautifully written; the text figures are well selected (those showing the serial sections of the rostral ends of brachiopods being particularly useful); the Index is serviceable; and, finally, the many plates are very well done, though unfortunately the figures on Plate 36 are reversed.

CAREY CRONEIS

Grundwasserkunde. By W. KOEHNE. Stuttgart: E. Schweizerbart'sche Verlagsbuchhandlung, 1928. Pp. 291; figs. 100. $4.50.

This is a very useful textbook on the geology of ground water, giving the present status of this science in a condensed form, and illustrated with sufficient formulas to be of value for the engineer and economic geologist. The book contains a good bibliography and a dictionary of all technical terms used in ground-water geology. It is well for geologists in English-speaking countries to keep in mind where they can find German technical terms explained as they are in this work on ground-water geology.

A. C. NOÉ

Streifzüge eines Geologen im Gebiet der Goajire-Indianer. Kolumbien. By OTTO STUTZER. Berlin: Dietrich Reimer, 1927. Pp. 154; pls. 32; chart 1; map 1. M. 10.

Otto Stutzer, who is now the director of the Institute for Fuel Geology in Freiberg, Saxony, and who also holds a professorship in the famous school of mines at Freiberg, spent two years in the republic of Colombia as a petroleum geologist, visiting in 1925 the Goajira peninsula at the request of the government of Colombia. A by-product of this work is this geologic travel-book which is an interesting mixture of geologic, topographic, and ethnographic observations told in popular language. The excellent illustrations contained on thirty-two plates, as well as the map, add to the value of the book.

A. C. NOÉ

Elektrische Bodenforschung. By W. HEINE. *Sammlung geophysikalischer Schriften* herausgegeben von C. MAINKA. No. 8. Berlin: Gebrüder Borntraeger, 1928. Pp. xi+222; figs. 117.

The author of this book describes the methods of geophysics used for the discovery of minerals. His years of practical application and theoretical study enable him to discuss his subject in a critical way. He discusses only briefly the geologic side, giving most attention to the methods of applied physics. The author tries to convince the reader that the application of geophysical methods to the study of geological problems can only be successful where the co-operation of the geologist and physicist is assured, and that neither one nor the other alone can give satisfactory results. The book is divided into three subdivisions which deal with, first, the measuring of the electric field; second, the magnetic field of electric currents in the soil; and third, the application of electromagnetic waves to the discovery of conductive deposits. No complete bibliography of the subject is attempted but references are made to the most important publications.

A. C. NOÉ

An Introduction to the Study of Ore Deposits. By F. H. HATCH. London: George Allen & Unwin, Ltd., 1929. Pp. 117; figs. 31.

Drawing upon his rich and varied experience in the fields of geology and engineering, Dr. Hatch has, in this little book, summarized in very acceptable fashion the modern conceptions of the methods by which ore deposits have been formed. For the mining engineer who wishes to familiarize himself briefly with the prevailing theories of ore formation, or for the student of other branches of geology who wishes a general view of the principles of ore deposition, the book should prove interesting and valuable. It can be easily read in one or two evenings.

E. S. B.